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THE TECTONIC HISTORY OF NEW ZEALAND

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Summary

The tectonic history of New Zealand falls into two periods of major activity, a late Mesozoic–early Tertiary phase, and a late Tertiary–Quaternary phase. The first phase is here termed the Rangitata Orogeny and the last phase the Kaikoura Diastrophism. The Rangitata Orogeny started in the southern part of the South Island probably early in the Jurassic and progressed northward bringing more and more of the New Zealand Geanticline above sea-level. The nose of the geanticline reached the North Island area late in the Cretaceous and the geanticline was above sea-level for the whole of the New Zealand area early in the Tertiary. There is evidence that the New Zealand Geanticline rose above sea-level earlier in the northern part of the North Island of New Zealand, but this is at present interpreted as a local undulation in the geanticlinal body. On the eastern and western sides of the geanticline, secondary, sedimentary basins—the Eastern Basin and the Western Basin—developed, in which most of the Cretaceo–Tertiary sedimentation in New Zealand took place.

While the geanticline was being brought above sea-level from the New Zealand Geosyncline, a force acting from north-north-east to south-south-west deformed the geanticline and pressed it westward into an arc—the Alpine Schist Arc. This internal deformation caused the development of the schistose rocks in this arc by stretching and slip in the Southern Alps part of the arc, and by crumpling and a minor amount of stretching in the Otago part of the arc. The stretching in the Southern Alps region probably reached its maximum in the late Cretaceous, and thereafter no schistose rocks were formed.

During the westward movement of the Alpine Schist Arc, the Western Basin was overwhelmed in the area between the arc and the Fiordland–Stewart Island Complex. The hypothesis that the Alpine Schist Arc is a large nappe is herewith proposed. Ultrabasic rocks are in such a position that

N.Z. J. Geol. Geophys. 2 : 1-55.

they could very well be situated on the thrust plane of the nappe. Gravity anomalies are most likely to have been caused by the overfold. The Alpine Schist Arc was broken up after the Eocene, because, at that time, the deforming movements were active in a more east-west direction. Immediately prior to this external deformation, the Alpine Arc had been wrenched off the northern part of New Zealand Geanticline in the Marlborough Sounds area, thus causing the different attitudes in the schistosities north and south of the Wairau Fault.

The breaking-up continued until the present, but the speed of deformation increased markedly during the Kaikoura Diastrophism. During the latest stages of the internal deformation and the earliest stages of the external deformation, faults showing dextral transcurrent movement developed parallel to the geanticlinal ridge. These faults swung westward during the Wairau wrench-off and continued across the geanticlinal body, cutting it into compartments during the external deformation. These faults have cut the Marlborough–East Coast area into a large number of parallel segments. As the faults are all dextral transcurrent, all the segments must move in a southerly direction. One of these segments, the Ruahine–Rimutaka Horst, was uplifted during the development of the Wanganui–Petane Trough; its continuation in the Marlborough area is the Spenser Mountain segment between the Wairau and Awatere faults.

INTRODUCTION

Field work in the North Island, and particularly on the North Island main ranges, has led the writer to the conclusion that the theories hitherto advanced on the structure of New Zealand cannot account for the assembled evidence. Not one of these theories is able to explain the large number and variety of facts that are now available for the interpretation of the tectonic history of New Zealand. None is able to explain the gravity anomalies, the regional distribution of metamorphic rocks, or the regional distribution of transcurrent faults.

Much of the evidence on which the present theory is based is available in published papers and in reports held by the New Zealand Geological Survey. Many years of personal field observations and re-investigation of the evidence have led to conclusions presented in this account. Observations and deductions not specifically attributed to other workers are those of the author.

The New Zealand Tectogene is comparable with the Swiss Alpine and Himalayan Tectogenes in size, but, as is to be expected, its structure has developed along completely different and independent lines. From field work in the Ruahine Range and from observations made in the Plio–Pleistocene geology of Hawke's Bay and Wanganui, the writer was led to conclude that the Kaikoura Diastrophism was exceedingly severe, and must have had a profound influence not only on the North Island but also on the South Island.

PREVIOUS THEORIES

Concerning the geological structure of New Zealand, and particularly the New Zealand Alps, a large number of geologists have expressed opinions.

Von Hochstetter (1863) was of the opinion that in the New Zealand Alps only the eastern half of a complete mountain chain had been preserved. In 1864, he explained the development of Cook Strait and Foveaux Strait by extensive block movements, which displaced the islands sufficiently to cause the ruptures.

Von Haast (1879) and Hutton (1900) supported Hochstetter's view and considered the Alps to be the eastern wing of an anticlinal arrangement of which the western wing was either submerged below the ocean or had been destroyed by erosion.

Marshall (1905) and Park (1910) accepted Hochstetter's view without further comment.

Suess (1906) accepted the idea that two independent "unilateral" mountain chains converged in the southern part of the South Island, forming a syntaxis.

Morgan (1908, 1911) had observed that the alpine schists are overturned on the western side of the Alps and that the faults on that side are of the overthrust type. His explanation for the structure of the Alps was, therefore, that the Alps had been thrust westward against immovable "buttons" of the Greenland Series. Because of the profound differences in attitude and metamorphism between the Greenland Series and the Alpine Schists, he could not accept the Greenland Series as a part of the core of Hochstetter's Anticline.

Van Bemmelen (1933) considered New Zealand to be a median geanticline of a large geosyncline. The geosyncline had been folded by a late Cretaceous orogeny, whereas Australia had been subjected to early Tertiary uplift. In a compensatory movement, the Tasman Sea and the New Caledonia Trough subsided, and a median ridge, the Lord Howe Rise, was raised in the course of subcrustal differentiation processes (Undation Theory).

King (1939, 1940) was of the opinion that deformation has affected the entire New Zealand region, that both islands were moving to the west, but that the South Island had outstripped the North, causing dislocation at Cook Strait.

Macpherson (1946) accepted the circum-pacific seismic belt as running through the central and southern portions of New Zealand. Broadening northward, the belt occupies the Fiji Basin and reaches eastward as far as the Tonga–Kermadec Trench.

Gutenberg and Richter (1949) pointed out that lack of definite records of seismic activity in the Tasman Sea suggests that this area is essentially part of the stable Australian continental shield, bordered to the east by the active circum-pacific seismic belt, in which New Zealand is situated.

Lillie (1951), discussing the geological structure of New Zealand, pointed out that a north-west trend, most clearly seen in the Mesozoic strata, is generally attributed to an early Cretaceous orogeny. A north–east trend, imparted in a late Tertiary orogeny, is dominant in both structural plan and physiography. Both trends were at one stage attributed to the early Cretaceous and gave rise to Suess's concept of syntaxis. Areas are known showing a distinct swing from north-west- to north-east-trending folds. Suess's idea of syntaxis accounts for the fold plan in the Tertiary rocks just as well as does the concept of arcuate structure. The validity of both concepts depends on precise definition in terms of time. Parallelism of trends for consecutive fold movements, particularly in the Tertiary strata, occurs in many parts of New Zealand.

Brodie (1952) discussed the major features—Lord Howe Rise, New Caledonia Trough, Norfolk Ridge—which all exhibit an arcuate northwesterly trend away from New Zealand. Assuming that the Lord Howe Rise is anticlinal in structure, he considered the continuation of this rise on the eastern side of New Zealand to be offset some 300 to 400 miles at the Kermadec Trough, the continuation being found in the Campbell Plateau. Accepting an earlier south-south-westerly directed thrust as the cause of the Tasman Sea ridges and basins, and of the later displacement of the Chatham Rise–Bounty Basin–Campbell Plateau in relation to the Norfolk Ridge–New Caledonia Trough–Lord Howe Rise, he assumes a subsequent eastward swing in thrust direction to explain the north-east-trending Tertiary fold systems of New Zealand.

Cotton (1956), discussing the geomechanics of New Zealand mountain building, concluded that a "geosuture" (Cloos, 1948) or a "regmatic" joint (Sonder, 1947), or a "transcurrent fault zone" (Wellman, 1955b) is the mean feature of New Zealand tectonics. Stresses set up by lateral drift in this zone produced block faulting and orogeny.

Wellman (1956), though presenting no general treatment of the structure of New Zealand in his treatise under this title, discusses sedimentation conditions in the various districts of the country during the geological past. A movement of 300 miles along the Alpine Fault is suggested. The movement is considered to have separated Nelson from Otago, on the basis that similar stratigraphical columns can be drawn up for both areas and that active dextral movement can be proved along the Alpine Fault today. The Alpine Fault is considered to continue northward through the volcanic region of the North Island.

NOTES ON THE STRUCTURE OF THE NORTH ISLAND

THE NEW ZEALAND GEANTICLINE IN THE NORTH ISLAND

Outline of the Geanticline and the pre-Pliocene Orogeny

Data on the movements of the North Island Geanticline have been collected over a long period, but nowhere has the sequence of events been more fully re-traced than in Hawke's Bay (Kingma, 1957b). In the Hawke's Bay sector, the North Island Geanticline started to rise late in the Cretaceous, and continued to rise until the Miocene. The geanticlinal crest was eroded continuously during its uplift and probably no extensive mountain ranges were formed. The generally coarse facies in the Tertiary strata of the Wairarapa suggest that the geanticlinal crest was higher above sea-level in the southern part of the Wellington Peninsula than in the north. Apart from slight local differences in orogenic movement, the geanticline thus rose faster in the southern than in the norther part of the Wellington Peninsula, consequently forming a higher relief in the south. ,

Still farther north, beyond the Wellington Peninsula, some very coarse sediments in the Tertiary sequence suggest that in this area also the North Island Geanticline was higher above sea-level. Hornblende gneisses and hornblende quartz diorites (P. 21971, Mr A. Steiner, pers. comm.) in the lower Ihungia Formation (Altonian stage) indicate that somewhere in the Rotorua-Coromandel area the axial portion of the geanticline was being eroded. These igneous conglomerates are preceded by tuff bands in the immediately underlying beds, and it is most likely that volcanic activity preceded the production of these conglomerates. Throughout the Tertiary, the geanticline appears to have been least elevated in the Hawke's Bay sector, and, as will be shown later, this is an important factor. The facies in the lower Cretaceous of Poverty Bay (Fig. 1) coarsens in a westerly direction, as also do igneous con-conglomerates, and it is most likely that at this time a part of the North Island Geanticline was above sea-level somewhere in the Rotorua-Coromandel area or perhaps farther north.

Whatever the irregularities involved in the rise of the North Island Geanticline, the geanticlinal movements belonged to one and the same orogeny, closely associated with the orogeny in the South Island (the post-Hokonui Orogeny), and is therefore here identified with it. The post-Hokonui Orogeny was thus not confined to the South Island, but embraced the entire New Zealand region. As the term "post-Hokonui" is rather misleading in that it could be taken to imply a "Hokonui" orogeny, which has never been postulated, the term post-Hokonui Orogeny is here abandoned in favour of "Rangitata Orogeny", indicating a New Zealand-wide orogeny, and leaving the term "Hokonui" exclusively for stratigraphical and lithological purposes. The name Rangitata was first used by Park in 1921, when he recognized five orogenies of which the Rangitatan was a "post-Hokonuian", early Cretaceous folding accompanied by ultrabasic intrusions.

The crest of the geanticline in the North Island was roughly 100 miles wide; the axis is recognizable in the Marlborough Sounds area where schists come to the surface, and in the National Park area where schistose, granodioritic, and gneissic rocks have been brought up by effusiva (Steiner, 1958). The drowned topography of the Marlborough Sounds indicates that the core of the geanticline plunges northward underneath the Wanganui depression, and there is no need to postulate any alteration in direction for the geanticlinal core north of the Sounds. The core region of the geanticline comes to the surface in the Permian rocks of Northland (Hornibrook, 1954) and the strata of Coromandel are probably close to it (Figs 1, 2). A line connecting these points shows an almost perfect arc (Fig. 1). The regional strike swing in the southern part of the Wellington Peninsula changes from north-west (030°) in

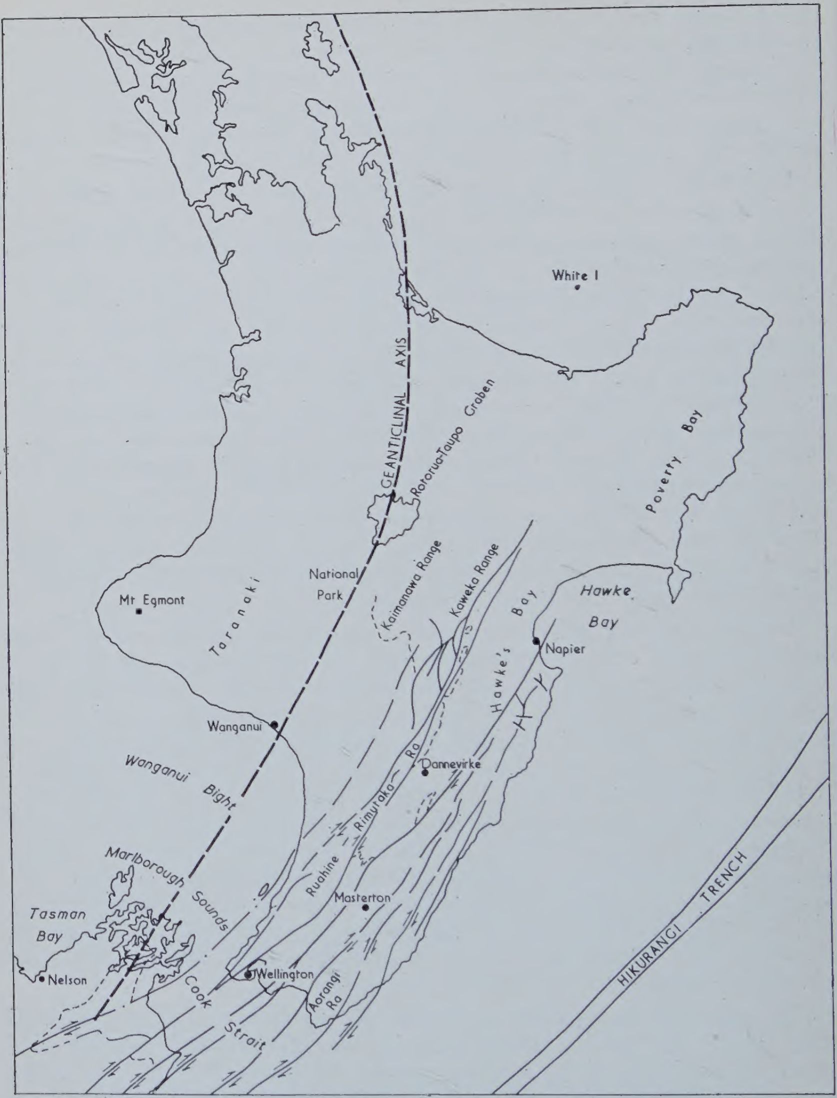

FIG. 1.—Outline of North Island, showing main topographical features, transcurrent fault zones in the Wellington Peninsula, and location of geanticlinal axis.

the Wellington area (Reed, 1957; Mr H. E. Fyfe, pers.comm.) to almost north-south in the central part of the North Island. In the Rimutuka tunnel, a constant 030° strike has been observed (Reed, 1957). The width of the crest of the geanticline has been taken as the distance separating the line on the east coast where the geosynclinal

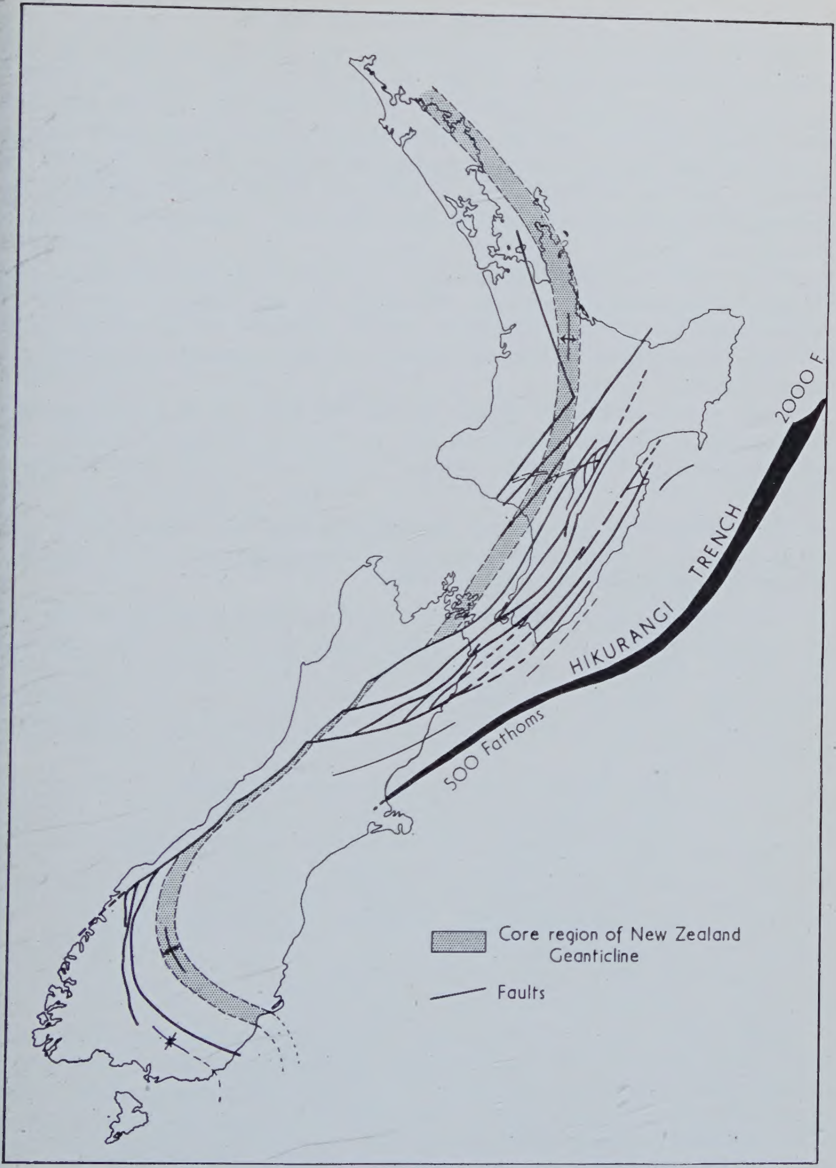

FIG. 2.—Outline of New Zealand, showing main fault systems and core of the New Zealand Geanticline.

Jurassic rocks crop out and the line on the west coast, where either the Jurassic rocks change from shelf facies to geosynclinal facies, or where the basement rocks are covered by shallow, Tertiary sediments, indicating the vicinity of a shore line.

The Ruahine-Rimutaka Horst

The Ruahine-Rimutaka ranges form the mountain chain in the southern part of the North Island. The rocks consist mainly of graded beds of Jurassic-Triassic age. These graded sequences can be followed over great distances and have been given various names (Waipapa Series, Alpine facies, Aorangi Series, Richmond Series, etc.). There seems to be little doubt that these beds are widely spread and may be encountered over the whole of New Zealand, ranging in age from Permian to Jurassic. It is here proposed to name these geosynclinal sequences the Aotearoa Geosynclinal Sequence, as distinct from the Hokonui Sedimentary System, replacing the confusing term "Alpine Facies", which refers particularly to the eastern facies of the New Zealand Geosyncline. It may in future also be necessary to change the name Hokonui as well.

In the easternmost part of the Horst, the finest parts of the graded cycles contain Jurassic pollen, on which Dr R. A. Couper, of the New Zealand Geological Survey, reports as follows (see also Kingma, 1958d):

"Two samples of alternating fine-grained sandstone and muddy siltstone from the Tutaekuri River, near Kuripapanga (Grid. ref. N. 123/825 574) were collected early in 1958.

"The sediments are broken only with difficulty in the hand, but are less indurated than the typical main range greywacke. The samples N. 123/516 (L 1066), and N. 123/517 (L 1060)—which is 50 yards downstream from 516—yielded only poor floras. Of the 21 species determined, only the following are known as yet to have stratigraphic value:

Classopollis torosus (Reissinger) Couper
Pteruchipollenites cf. *thomasii* Couper
Caytonipollenites pallidus (Reissinger) Couper
Microcachryidites antarcticus Cookson
Zonalopollenites cf. *segmentatus* Balme
cf. *Baculatisporites truncatus* (Cookson) Balme
Osmundacidites n.sp.
Trilites n.sp.
Polypodiidites n.sp.

"None of the species listed above is known with certainty from beds older than Jurassic. *Microcachryidites antarcticus* is known only from middle Jurassic and younger beds in Australia and New Zealand, while at present the three new species noted above are known elsewhere only from middle Jurassic (Temaikan) beds at Kawhia. An age no older than middle Jurassic thus seems likely. The upper age limit is not easily assessed.

Zonalopollenites segmentatus is apparently confined to Jurassic beds in West Australia and has been noted in Jurassic beds in New Zealand. *Classopollis torosus, Caytonipollenites pallidus* (a member of the Caytoniales), and *Pteruchipollenites* cf. *thomasii* (a Pteridospermous pollen grain) are occasionally found in lower Cretaceous sediments, but are very characteristic of Jurassic floras in north-west Europe, Australia, and New Zealand. At the present stage of knowledge, plant microfossil evidence in the above-mentioned samples favours a middle to upper Jurassic age for the greywackes at Kuripapanga."

The Jurassic-Triassic beds are usually vertical, but anticlinorial and synclinorial axes have been mapped on evidence of top and bottom in the graded cycles. Cores of some of the anticlines in the northern Ruahine Range have been observed to pitch as steeply as 70° to 80° south.

Tectonically, the chain is a horst (Kingma, 1957c) in which Mesozoic rocks come to the surface. Both on the eastern and western sides, major faults separate the Mesozoic rocks of the horst from their equivalents on the east and west coasts. The maximum throw is of the order of 5,000 ft on the eastern and 6,000 ft on the western side of the horst. This 200-mile-long body is only a sliver when compared with the original width of the geanticline; its situation on the eastern flank of the geanticline is, from a tectonic point of view, rather unusual. The horst is crossed by a number of faults, cutting it into a series of en echelon fault blocks (Figs 1, 2, 3). These faults and the bordering

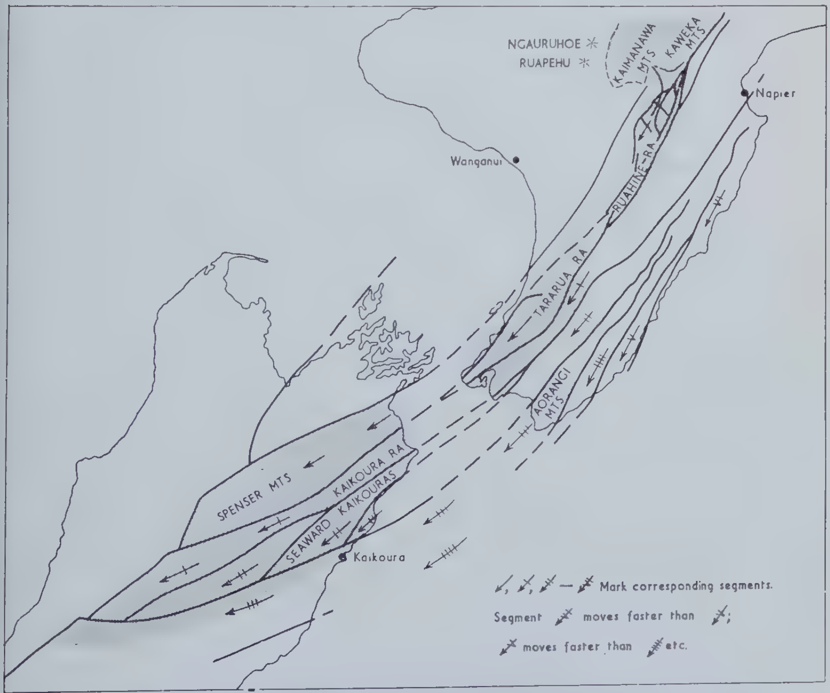

FIG. 3.—Outline of Wellington Peninsula and Marlborough Province, showing transcurrent fault zones and segments between these fault zones culminating in the Spenser Mountains, Inland and Seaward Kaikouras. To maintain dextral transcurrent movement on fault zones, the more easterly segments move faster towards the south.

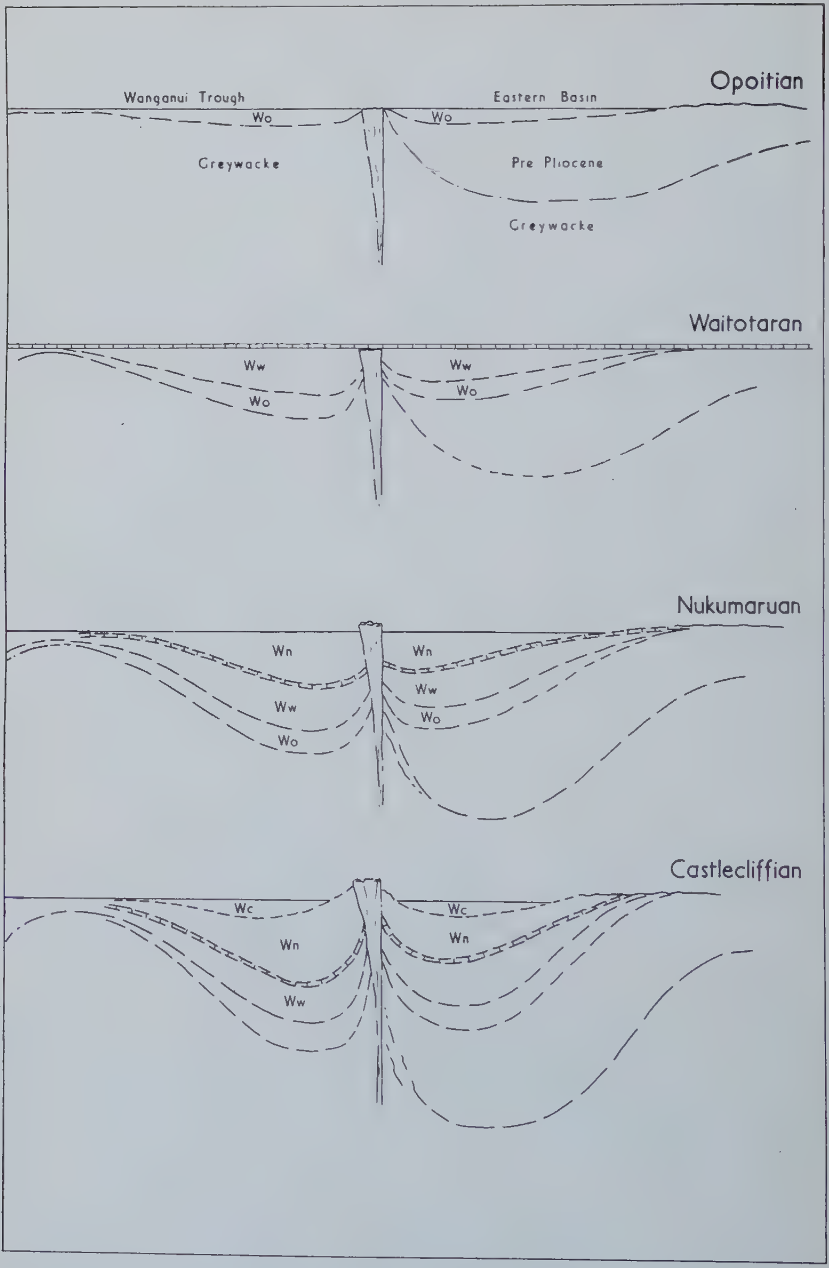

FIG. 4.—Sketch of sedimentation in the Eastern Basin and in the Wanganui Trough during the Plio-Pleistocene, showing deformation of the pre-Pliocene basement and rise of the Ruahine-Rimutaka horst. (Vertical scale: one inch $\cong$ 5 miles.)

faults all have an appreciable clockwise lateral component. As all the faults are dextral transcurrent, a distinct southward movement of the en echelon blocks has taken place. The main faults may continue southward into the Wairau and Awatere fault zones of the Marlborough district (Figs 3, 11), while other transcurrent fault zones in Hawke's Bay and Wairarapa may be the continuations of the more southerly Marlborough fault zones. The horst is heavily fractured internally by the transcurrent and vertical movements, creating rapidly eroding, unstable rock faces.

It appears that while most of the geanticline in the southern part of the North Island was being submerged at the end of the Rangitata Orogeny, the narrow Rimutaka-Ruahine horst was forced up as an enormous piercement body (Kingma, 1957c). The faults bordering the horst on the western side are all of the reverse type and dip 70° to 80° east, while those on the eastern side are mostly vertical. If these fault planes extend to great depth, while maintaining the same dip—which, judging from the size of the horst and the surface expression of the faults, they probably do—the horst has the shape of a large asymmetrical wedge, with its apex probably terminating in a single crushzone approximately 5 to 10 miles deep (Fig. 4). No gravity anomaly is associated with the horst; the gravity pattern shows up as if the horst were simply not there (Fig. 6). This evidence strengthens the assumption that the base of the horst is a single crushzone and that this mountain chain has no real root.

The Kaimanawa Range

The Kaimanawa Range is situated immediately north of the Ruahine Horst, but occupies a much wider territory. Bordered on the western side by the Rotorua-Taupo Graben and on the eastern side by the Ruahine-Mohaka fault zones, these mountains represent the most complete flank of the North Island Geanticline (Fig. 1). Northward, the range continues as the Ahimanawa and Huiarau ranges, while on the southern side the range is faulted and the high relief gives place to the much lower relief of the Taruarau Saddle (Fig. 5).

In the easternmost part of these ranges, the geosynclinal graded beds have now been formally assigned to middle Jurassic (Kingma, 1958d, and Couper (see above)). The rank increases westward until in the westernmost part slightly deformed greywackes and greywacke conglomerates (Mr G. W. Grindley, pers. comm.) predominate. The structure of these ranges is complicated, but probably best considered as an anticlinorial system in which anticlinal axes plunge very steeply north and south. This plunge probably represents a superimposed folding resulting from a change in direction of the principal horizontal stress acting on the geanticline after the anticlinorial system had been established. No major faults parallel to the strike are known to cross the ranges.

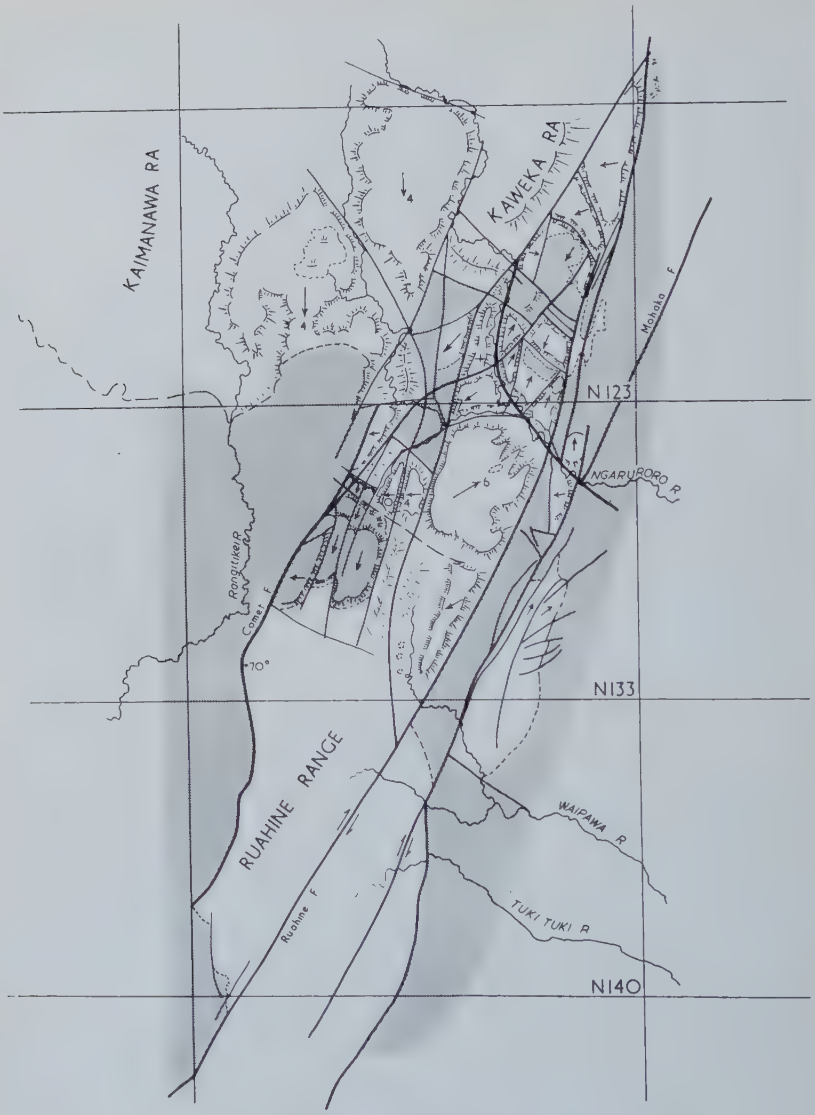

FIG. 5.—Fault systems in Sheets N 123, N 133, and N 140, showing extreme case of block faulting. Arrows indicate direction in which fault blocks are tilted. Shaded areas denote Tertiary strata. The depressed area between the Ruahine and Kaweka ranges is the Taruarau Saddle.

The Taruarau Saddle

The relatively narrow area between the Ruahine horst and the Kaimanawa Range is strongly faulted (Fig. 5) and infaulting of Mio-Pliocene rocks over thousands of feet has caused a spectacular sag in the main divide. This sag is also the location where the Rangitikei-Waiapu Anomaly (Robertson and Reilly, 1958) crosses the island from Wanganui to northern Hawke's Bay (Fig. 6). The movements of the fault blocks can be traced from the attitudes of the Kaukau surface in this area (Kingma, 1958b); it is therefore highly probable that the Ruahine horst was torn off the Kaimanawa Range, leaving the Taruarau Saddle, or graben, behind as the movement progressed.

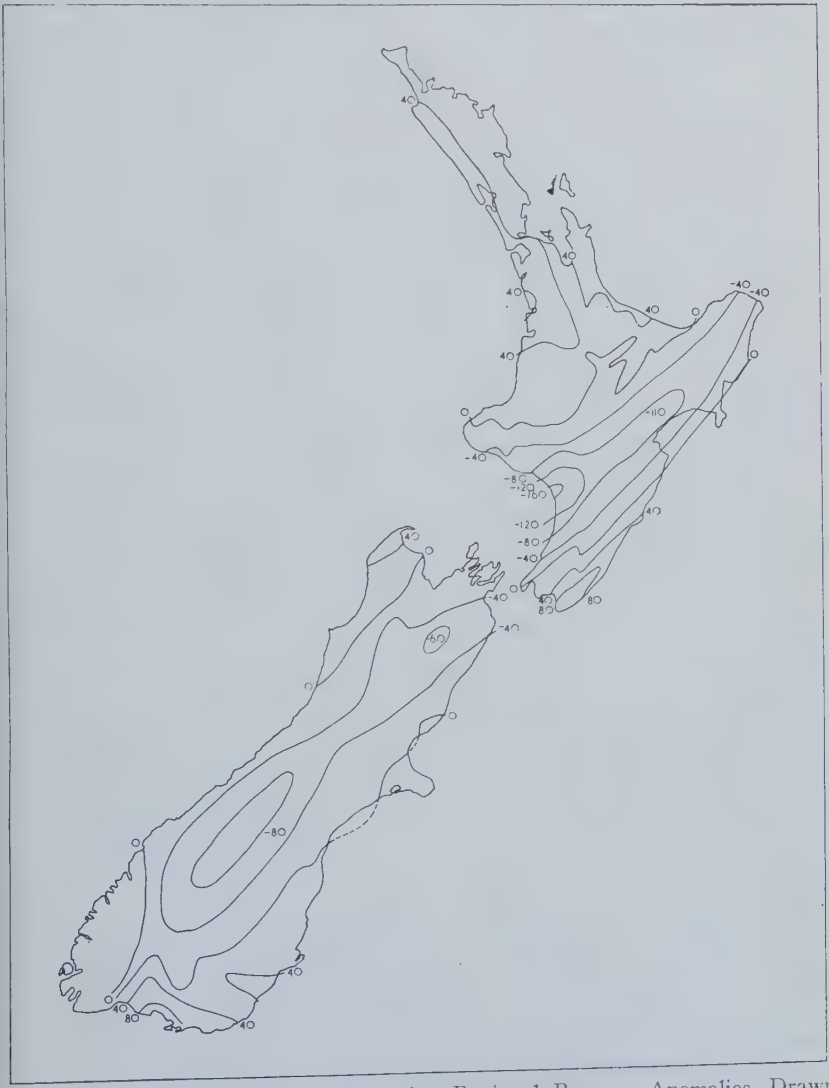

FIG. 6.—Outline of New Zeland, showing Regional Bouguer Anomalies. Drawn from data by Geophysical Survey, D.S.I.R., after Robertson and Reilly (1958). (Bouguer anomaly values in milligalls.)

THE PRE-PLIOCENE SEDIMENTARY BASINS EAST AND WEST OF THE NEW ZEALAND GEANTICLINE IN THE NORTH ISLAND

From general observations made on the eastern and western sides of the North Island, it may be concluded that sedimentation during Cretaceo-Tertiary times took place east and west of the New Zealand Geanticline. Field work by Ongley (1935), Lillie (1953), Kingma (1957b), and Wellman (1956) in Hawke's Bay and in the Wairarapa has shown that sedimentation has been continuous during the Cretaceo-Tertiary and that the unconformities that occur are not of regional significance and represent only brief time intervals. The sedimentation took place in what Macpherson named the Eastern Geosyncline (Macpherson, 1946; Wellman, 1956; Kingma, 1957b). This basin of deposition was called a geosyncline by Macpherson; for convenience, this feature is here termed the Eastern Basin. This Eastern Basin, and the sedimentation that took place therein, can be traced southward from the east coast of the North Island into Marlborough and Canterbury. The lithologies and facies changes are so readily traceable that the conclusion is inescapable that the Eastern Basin extended from South Canterbury northwards as far as Poverty Bay. It probably extended even farther, for in Northland a lithology very similar to that of the east coast of the North Island has been found (Mr R. F. Hay, pers. comm.).

In Marlborough, the sequence has been disturbed by faulting and the emergence of the Kaikoura blocks, and in the southern Wairarapa the Aorangi Range is a very late feature, causing unconformities in the lithological sequences. In Canterbury, Hawke's Bay, and Poverty Bay, the stratigraphy clearly demonstrates the unity of the basin. The Eastern Basin has been marine throughout, except in South Canterbury, where coal measures are found at the base of the sequence and terrestrial conditions prevailed at least in the lowest part of the Tertiary and the Cretaceous.

On the western side of the geosyncline, a Western Basin stretches from Westland (Gage, 1949, 1952) through Nelson into Taranaki and farther north into the Waikato and Auckland areas. This basin was named the Western Geosyncline by Macpherson (1946); the beds are marine, although more coal-bearing deposits occur than in the Eastern Basin. In the western and southern parts of the South Island marine Cretaceous beds are entirely replaced by freshwater beds that include many coal-bearing sequences. The basin is at present interrupted by the Tasman Bay-Wanganui Bight, which is mainly a Pliocene and post-Pliocene feature. Such important deposits as the Oligocene Te Kuiti Limestone can be correlated with the Cobden Limestone, in Westland, and there seems to be little doubt that the pre-Pliocene rock sequences can be well correlated over the entire Western Basin (Dr J. van der Sijp., pers. comm.).

The evidence therefore requires recognition of two basins—an Eastern Basin and a Western Basin—in both of which marine sedimentation took place and between which the New Zealand Geanticline acted as a barrier. This barrier was cut down continuously while the Rangitata Orogeny progressed, supplying sediment for both basins. The metamorphic rocks of the Southern Alps were apparently not eroded until the Pliocene, as the first appearance of schist pebbles is in the Old Man Gravels of Westland. At times the barrier was so low that sedimentation in the two basins merged and small infaulted inliers (Macpherson, 1949) remain as evidence of this. No generalizations about the sedimentation can be made, except perhaps that bentonitic muds are more widespread in the Eastern than in the Western Basin and that the Pareora sedimentation in the Western Basin resulted in a greater thickness than in the Eastern.

THE PLIO-PLEISTOCENE SEDIMENTATION EAST AND WEST OF THE RUAHINE-RIMUTAKA HORST

The sedimentation in the Eastern and Western basins during the Plio-Pleistocene did not follow the general pre-Pliocene trend; during this period, the New Zealand Geanticline was suppressed in the area between the Marlborough Sounds and National Park (Figs 1, 2), and the wedge-shaped Ruahine-Rimutaka horst acted as a barrier between the Eastern and Western basins. The Eastern Basin was less affected, but the Western Basin spread eastward across the crest of the submerging geanticline. Movement along trancurrent fault zones on the east coast continued during this period, coming to a culmination during the Kaikoura Diastrophism.

On the eastern and western sides of the Ruahine-Rimutaka horst, deep, marine sedimentary basins developed during the Plio-Pleistocene in which huge quantities of sediment accumulated. Elsewhere in the Eastern and Western basins, sedimentation was slow, being terrestrial over large areas and, except locally in fault angle depressions, was not of great thickness. Contrary to favoured geological concepts, these basins increase in depth towards the tectonic high, accentuating again the extraordinary shape and tectonic position of the Ruahine-Rimutaka horst. Only those parts of the Eastern and Western basins located immediately east and west of the Ruahine-Rimutaka horst need further explanation here. They are the Petane Trough and its extension into the Wairarapa on the eastern side, and the Wanganui Trough on the western side of the horst.

The Petane Trough

The axis of the Eastern Basin shifted from time to time during the pre-Pliocene, but it was not until the Pliocene that a pronounced westward shift was initiated which continued during the Plio-Pleistocene, developing a deep basin immediately east of the rising Ruahine-Rimutaka horst (Fig. 4) (Kingma, 1957c). The movements in this basin during the Pliocene were mild in comparison with those in the Pleistocene, when the Kaikoura Diastrophism came into full force. The outline of this basin during lower Pleistocene, Nukumaruan time, and the thicknesses of sediments (Kingma, 1955) therein are indicative of the movements then taking place. That part of the basin where most sediment was accumulated and in which 10,000 ft of Nukumaruan sediment has been measured (pers. observ.) is situated immediately north-west of Napier, and has been named the Petane Trough (Fig. 7). The sequence in this trough dies out abruptly northward towards Wairoa, on the northern shore of Hawke Bay. From the Petane Trough, a shallower basin continues southward, flanking the Ruahine-Rimutaka horst, as far as Cook Strait. There seems to be no doubt that the development of the Hawke Bay as a tectonic feature—it being situated in the eastern part of the Petane Trough—is closely related to this depression, and that the Hawke Bay is the remnant of the Pleistocene Petane Trough.

The sediments for the Plio-Pleistocene deposition were derived mainly from the Ruahine-Rimutaka horst or from the Kaimanawa-Kaweka ranges, only a small quantity being supplied by low-lying country in the extreme east of Hawke's Bay and Wairarapa districts. The general trend and thicknesses of the Nukumaruan sedimentation are shown in Fig. 7. The Castlecliffian stage followed the same sedimentary trend, but did not become so thick as the Nukumaruan, due to the fact that the Petane Trough became very shallow and the above-mentioned lowland in the extreme east of Hawke's Bay and Wairarapa continued to emerge. From aggradation of rivers in the area immediately flanking the horst, it is concluded that terrestrial sedimentation is still in progress.

The Wanganui Trough

The Wanganui Trough, situated on the western side of the Ruahine-Rimutaka horst, exhibits many features similar to those observed in the Petane Trough. A thickness of 16,000 ft for all the Pliocene and Pleistocene stages has been measured, not only in surface exposures but also in drill holes.* The Pliocene rests immediately on greywacke basement, giving a difference in height of more than 20,000 ft between the greywacke surface in the Wanganui Basin and the abrasion surface on the Ruahine-Rimutaka horst. The sequence in the basin becomes thinner as the basement rises towards the west. The culmination of this rise is considered to be on a line roughly north-south through Patea, and from this basement high the entire Tertiary sequence thickens again westward. Thickening is greatest in the pre-Pliocene sediments, and the Plio-Pleistocene continues to thin. The thick Oligo-Miocene strata to the west were presumably deposited on the western flank of the North Island Geanticline.

The Wanganui Basin is intersected by numerous faults, some of which line up with the Rotorua-Taupo Graben, while others converge into the Taruarau Saddle (Fig. 2). The faults are almost certainly deep seated and cut through the basement rock, dividing it into a

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*Superior Oil Company, 1943, open file, New Zealand Geological Survey.

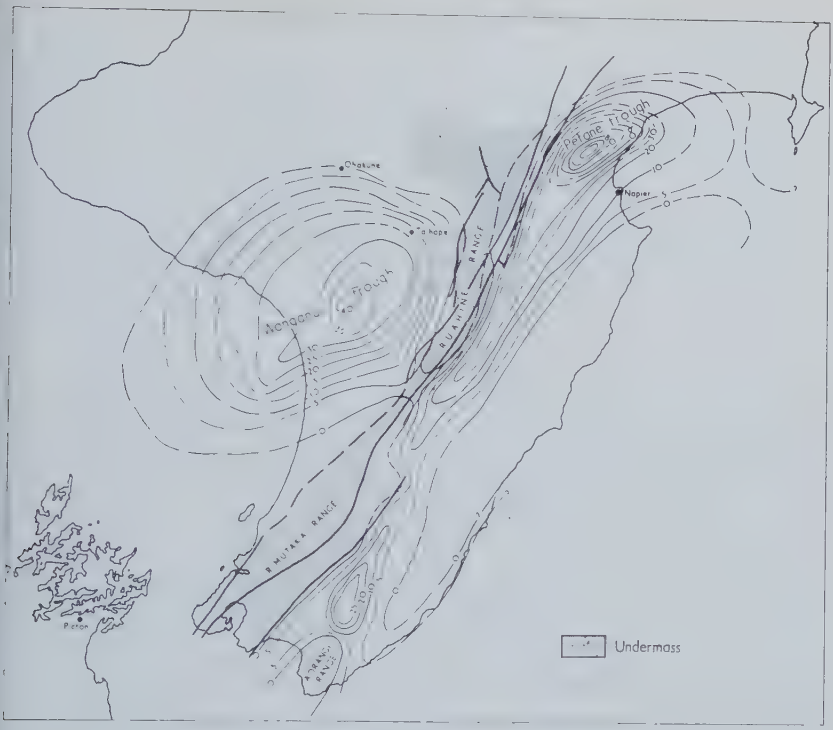


Fig. 7.—Isopach map of the Nukumaruan Stage with isopach intervals of 100 ft, showing extent of Nukumaruan sedimentation and consequently the deformation of the top of the Waitotaran stage (Kaikoura Diastrophism) in the Wellington Peninsula. On evidence of open-sea Mollusca in the Nukumaruan Strata of the Petane Trough, an opening to the ocean was very probably in existence in the Hawke Bay area, and the O-Isopach line has been drawn accordingly.

number of north-south-trending segments. Along the surface expressions of these faults, transcurrent movement has been detected; it appears that these segments are equivalent to those on the east coast, but the dextral transcurrent movement is much smaller.

In lower Pliocene (Opoitian) time, sedimentation in the Wanganui and East Coast basins was connected through the Kuripapanga Graben (Kingma, 1957b), but in upper Pliocene (Waitotaran) time the entire northern half of the horst submerged, allowing the Waitotaran sea to deposit coquina on the crest. The thickness of Pliocene sediments is rather small compared with that of the Pleistocene (Nukumaruan and Castlecliffian). The great thickness observed in Pleistocene sediments on both sides of the horst indicates that the main downward movements in these basins took place at that time. Probably the formation of these basins was a consequent result of the relatively speedy rise of the horst (Fig. 4).

THE ROTORUA-TAUPO GRABEN

The Rotorua-Taupo Graben is one of the major Tertiary features of the North Island. Its tectonic position is exceptional, as it cuts through the geanticline at an oblique angle. The fact that it cuts through the New Zealand Geanticline and the presence in and surrounding the graben of enormous quantities of acid effusiva suggest that the granite core of the geanticline was fused by basaltic magma. The importance of the Rotorua-Taupo Graben for the interpretation of the geology of New Zealand is emphasized; it is hoped that the vast quantity of geological and geophysical data collected over the past ten years will be published soon.

Development of the Graben

Kapitean sediments in Taranaki and in the Taihape-Taruarau area have not been contaminated by volcanic products, which leads to the assumption that no volcanicity was taking place in that area at the time. During the following Opoitian time, there are no indications that volcanicity was developing; Opoitian rocks contain normal erosion products from a greywacke body, presumably the North Island Geanticline (Fleming and Steiner, 1951). From the distribution of the Opoitian rocks, it is obvious that the Taruarau Saddle was well advanced in its development and provided a waterway (Kuripapanga Strait, Kingma, 1957a) between the eastern and western sides of the North Island. In upper Waitotaran time, the first rhyolitic products became incorporated in the marine sediments of the Taruarau Saddle, indicating that the explosive phases of the Rotorua-Taupo volcanism had started (Kingma, 1957a). The full impact of this volcanism was not evident until late Nukumaruan-Castlecliffian time, when rhyolitic tuffs were spread in huge quantities; the volume of volcanic dust in a single pumice band is assessed at several cubic miles. The present shape of the graben apparently started to develop in Waitotaran time, probably earlier, and was fully developed in late Castlecliffian time.

The Volcanics

Steiner (1958) has investigated the 1954 Ngauruhoe lava and its xenoliths. His conclusions were that volcanicity in the National Park area started with basic lava which became mildly contaminated with material from the geanticline, forming a basaltic-dacitic sequence. During the development of this sequence, a separate, acid magma was being formed from transfusion of acid gneiss in prolonged contact with an olivine basaltic magma at great depth. This acid magma produced the acid tuffs and ignimbrites of the Rotorua-Taupo Graben. The graben developed simultaneously with the blow-out.

Basement

Geophysical surveys along the fringes of the Rotorua-Taupo area have revealed that greywacke rock is above sea-level in large areas, but overlain by a thin cover of ignimbrite and pumice (E. I. Robertson, *in* Grange, 1955; F. E. Studt, pers. comm.). The thickness of the volcanic strata in the graben is unknown, but is likely to be between 5,000 and at least 10,000 ft, probably 15,000 ft or even more. Data on the basement are therefore few, and it is only because gneisses, granites, etc., have been brought up by lavas that it is possible to postulate that the core of the North Island Geanticline lies buried under the Rotorua-Taupo Graben. The presence of biotite quartz diorites in Coromandel, biotite granodiorites at Atiamuri (P. 15778), and granite boulders at Huka Falls (P. 15443) supports this assumption. The xenoliths in the Ngauruhoe lava described by Steiner (1958) point to the presence of acid gneisses underneath National Park.

The question has arisen as to why there is no evidence of volcanic activity along the deep-seated fault zones flanking the Ruahine-Rimutaka horst. This question cannot be answered with certainty. The faults are in the transcurrent-reverse category and consequently in a compressional area, and thus no outlet for volcanicity is available; the presence of a wide rift such as the Rotorua-Taupo Graben may have acted as an outlet for magma under the Wanganui depression.

PRINCIPAL HORIZONTAL STRESS DIRECTIONS

Following Lensen's principal horizontal stress chart for New Zealand (Lensen, 1958c) and taking the writer's hypothesis into account that transcurrent faults have been active in the East Coast area practically throughout the Tertiary (Kingma, 1958c), the established lineation of these transcurrent faults then becomes of utmost importance. According to Lensen's chart (Lensen, 1958c, fig. 2), this transcurrent lineation can be produced by principal horizontal stresses acting in any direction between 058° and 103°. This principal horizontal stress sector (P.H.S. sector) is presumably broadly similar for the entire New Zealand region.

In the North Auckland and Poverty Bay areas, a change in principal horizontal stress direction has taken place, as is evident from changes in fold axes as reported by Lillie (1953) and Brothers (1956), Dr H. J. Harrington (pers. comm.), and Mr M. C. Pick (pers. comm.); the earlier regional strike is north-west and the later north-east. The P.H.S. range for this strike swing is within the above-mentioned P.H.S. sector. When this P.H.S. sector is applied to the Hawke's Bay area, both the early and late Tertiary folding can be explained by a swing in P.H.S. direction; the transcurrent faulting is a by-product, and once initiated remains active in the lineation first established.

From personal observation, it is deduced that in lowest Tertiary times folding rather than transcurrent faulting took place, implying that the angle between P.H.S. direction and the later transcurrent fault lineation was greater than 56° (Fig. 8). Later in Oligocene time, when large amounts of transcurrent movement took place and the angle was approaching 45°, the transcurrent movement would have been at its peak. During this time the New Zealand Geanticline in the North Island was suppressed in the Wanganui area, thus allowing Landon

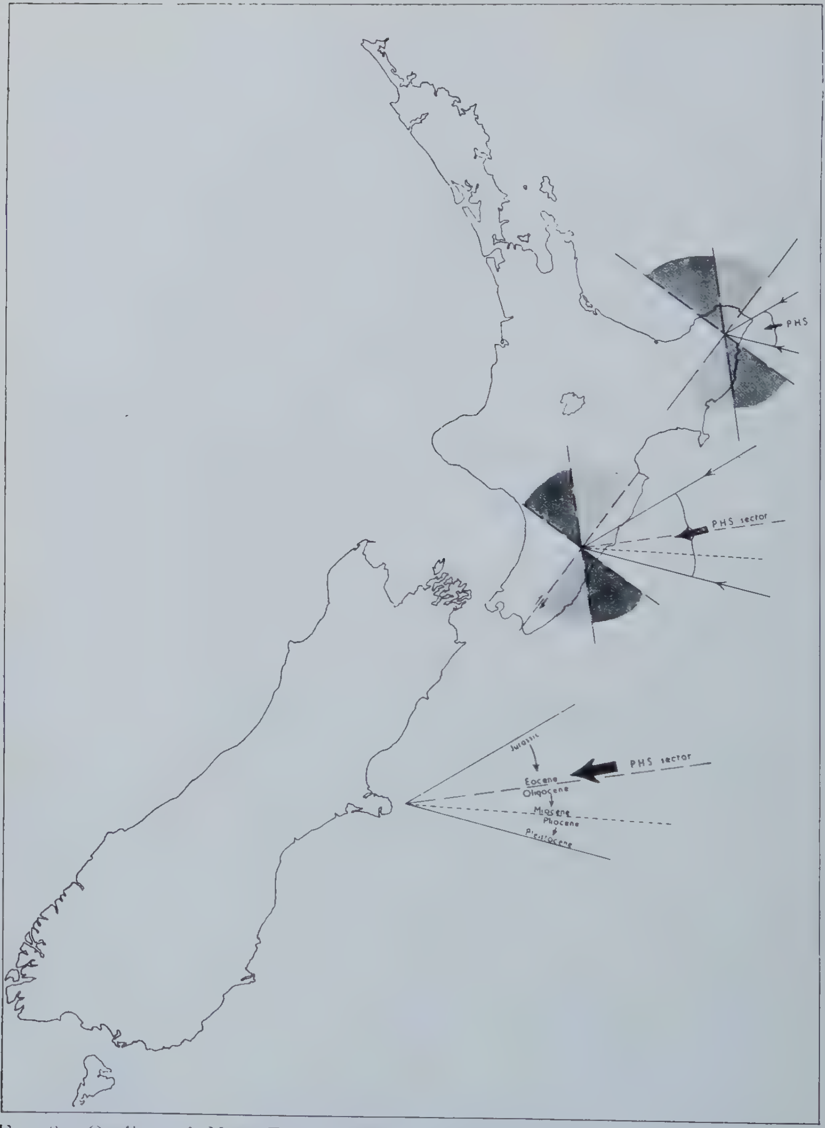


FIG. 8.—Outline of New Zealand and the influence of the Principal Horizontal Stresses. First stage P.H.S. direction is from north-north-east to south-south-west, resulting in fold axes in a northwest direction (Rangitata Orogeny). With change in direction, new fold axes were superimposed on the previous system (Kaikoura Diastrophism).

and Southland sedimentation to spread on to the New Zealand Geanticline. The detachment of the entire Wellington Peninsula from the remainder of the North Island became more and more pronounced as southward movement took place as part of the New Zealand Tectogene.

In Opoitian-Waitotaran time, a further swing in P.H.S. direction resulted in a greater degree of clockwise transcurrent-reverse faulting and caused the emergence of the Ruahine-Rimutaka horst (Kingma, 1957c). Movement, both vertical and transcurrent, affected this horst throughout the remainder of the Pliocene and Quaternary. The temporary submergence of the horst during the upper Waitotaran was probably the result of settling of this block after its separation from the Kaimanawa-Kaweka block. The North Island Bouguer anomaly, which has been named the Rangitikei-Waiapu Anomaly by Robertson and Reilly (1958) and which crosses the North Island from Wanganui to Napier, was caused by this movement and was apparently initiated in Oligocene time.

Lensen (1958c, fig. 5) has shown that the P.H.S. direction along the North Island transcurrent faults is uniform, but that there is an important swing of about 20° along the large faults in the Marlborough area. This means that, in late Quaternary time, the P.H.S. direction for this part of the South Island was more from the south-east than in the North Island. This difference in P.H.S. direction for the two islands was probably established earlier, and could account for the Wairau Wrench-off.

THE RANGITATA OROGENY IN THE NORTH ISLAND

From personal observations along the east coast of the North Island, and particularly in the area south of Hawke Bay, the movements of the pre-Kaikoura, Rangitata Orogeny have been deduced. The lower Cretaceous Taitai-Mokoiwi sedimentation in Poverty Bay derived its sediments from the west, and during this period the geanticline was above sea-level in the Rotorua-Coromandel area. The movements took place earlier in the area north of the east-west line through the Hawke Bay and lasted longer south of that line.

In the southern part, geosynclinal conditions prevailed in Upper Cretaceous time, and the geanticline emerged in the lowest Tertiary, but, judging from the fine sediments derived from it (Kingma, 1957b), the topography must have been low. Later in the Tertiary, the geanticline was above sea-level for the whole of the island, but the southern part was always lower, even to such an extent that Tertiary deposits transgressed across the geanticline from time to time. During the Kaikoura Diastrophism, the depressed part developed into the Wanganui Trough.

THE KAIKOURA DIASTROPHISM IN THE NORTH ISLAND

The severe deformation that affected the North Island during this cataclysm was particularly marked in the southern and central part of the island. While the central part was severely concomitant with these movements, there was considerable activity along the transcurrent fault zones in Hawke's Bay and Wairarapa, and the Kaikoura blocks emerged during this period. The original term for this deformation is "Kaikoura Orogeny", but as the term orogeny implies the formation of a mountain chain from a geosyncline, and as the Kaikoura Orogeny was really the deformation of an already existing geanticline, the term Kaikoura Diastrophism is here favoured. It is realized that deposits in small basins, marginal to the geanticline, were everted and severely folded during this activity and that therefore the term orogeny is correct for these areas. The Kaikoura mountain-building movements thus consisted of two categories, in which the deformation of the existing geanticline was much more important than the development of much smaller ranges from the marginal basins.

In the southern part of the island, the intensity of warping was greatest during the Nukumaruan and Castlecliffian. The movement continued through post-Castlecliffian time to the present, but the peak of the deformation has probably passed. The distribution of the Kapitean, Opoitian, and Waitotaran rocks indicates that movements had already started well before Nukumaruan time. These movements were at first only a crustal deformation of the Wellington Peninsula, but pressure due to the increasing intensity of the warping forced up one of the segments between two transcurrent fault zones; this segment is the wedge-shaped Ruahine-Rimutaka horst. The uplift of this wedge probably started in late Kapitean time, for in Opoitian time the Kuripapanga Graben was being filled with sediment, and, as may be judged from the Waitotaran sedimentation, uplift became more and more pronounced during the Wanganuian epoch. The facts, already mentioned, that the deposition does not thin out against the horst, and that, if the horst were removed, the Wanganui and Petane troughs would make one continuous, deep basin, shows very markedly the extraordinary position of the Ruahine-Rimutaka wedge. The assumption seems justified that the horst rose as a result of space shortage in the Wanganui-Petane downwarping, and of the southward trend in the transcurrent East Coast segments into Marlborough.

The pumice production of the Rotorua-Taupo volcanism, which, as Steiner (1958) has shown, belongs to the later phases of the volcanism, took place mainly in Pleistocene time, while the more basic volcanism was mainly a Pliocene feature. The fracture initiating the Rotorua-Taupo Graben and its northward extension, the White Island Trench (Fleming, 1952), was thus under development in early Pliocene time. During the Rangitata Orogeny, there was probably an early period of volcanicity in the area between Rotorua and Coromandel, due to the swing in horizontal stress direction by which the Northland area of the New Zealand Geanticline was pushed westward, causing a fracture in the Rotorua area. A rejuvenation of this movement in the Pliocene apparently took place.

The Poverty Bay area was severely affected by the Kaikoura Diastrophism in which a fold system with north-east axes was superimposed on an earlier system with north-west axes; no transcurrent fault-systems of any magnitude accompanied this deformation.

The North Island was thus influenced by two periods of deformation, in which the principal horizontal stress direction was different, the later period being much more cataclysmic than the earlier, more gentle Rangitata Orogeny. The geanticline was, by the end of the Miocene, subdued and eroded, and the orogeny had waned so, that, without the Kaikoura rejuvenation, New Zealand would scarcely have been above sea-level today.

NOTES ON THE STRUCTURE OF THE SOUTH ISLAND

THE SOUTHLAND SYNCLINE AND THE STRUCTURE OF THE EGLINGTON-HOLLYFORD VALLEY

Mutch (1956, figs 1, 2) outlined the stratigraphy and structure of the Southland Syncline. According to his summary: " . . . facies changes indicate the former existence of an old landmass beyond the southern margin of the Southland Syncline and a geosyncline beyond the northern margin. Two major, marginal unconformities are described. The north-east thickening of the Permo-Triassic beds is shown by isopach maps. The relation of the beds to the geosyncline and the Otago Schist is discussed . . . ." The rocks of the Southland Syncline may be traced as far north-west as the northern tip of Lake Te Anau, and Grindley (1958) has traced Permian sediments into the Eglington-Hollyford valley as far north-west as southern Westland, where the sequence is very much faulted, but where upper Permian beds could still be recognized. Dr J. B. Waterhouse (pers. comm.) has now found *Mytilus problematicus* in the Eglington Valley sediments, proving that at least the Carnian (Otamitan) is preserved there.

Wellman (1952) stressed the similarity between the Southland and Nelson stratigraphical columns, for the upper Paleozoic Matai and Te Anau groups both contain sediments deposited in a large geosyncline, with pillow lavas, extensive limestone, etc. This similarity, together with the fact that dextral transcurrent movement is evident today along the Alpine Fault, provided the evidence on which Wellman based his theory of a 300-mile displacement along this fault (Wellman, 1956).

Wood (1953) has shown that the north-eastern flank of the Southland Syncline is overturned, and Macpherson (1946) emphasized the overthrusting north-west of Balclutha, which can be traced over a distance of 50 miles. The Te Anau rocks in the Southland Syncline pass downward into the Otago Schists without any evidence of unconformity. The overall thickness of the combined Maitai and Te Anau sequences is of the order of 50,000 ft. They lie conformably between the schists and the Triassic sediments. Grindley (1958) has reported members of this rock sequence in fault blocks in the Livingstone Range (S. 131); he has mapped the Southland Syncline in that area as being tightly folded, overturned, and fractured betwen the Otago Schists and the pre-Carboniferous Fiordland-Stewart Island Complex. This complex is generally considered to be very much older than the rocks involved in the Southland Syncline and Otago Schists. The Greenland-Arahura-

Aorere-Mt. Arthur-Waiuta series of Westland are generally considered to be of pre-Carboniferous age also, and on this account are correlated with the Fiordland-Stewart Island Complex. As these rocks are very probably the remnant of a much larger mass that earlier must have extended into the present Tasman Sea and perhaps even farther south, this complex is here called the Tasman Cratogene.

The position of the Southland Syncline between the Otago Schists and the Fiordland-Stewart Island Complex has never been adequately explained. The thickness is 10 miles for the Te Anau-Maitai sequences combined; if the Southland Syncline were tightly folded, the total thickness of the combined flanks would be 20 miles. This presents the problem of how to accommodate this 20 miles of sediment in a belt approximately 3 to 5 miles wide in the Eglington-Hollyford valley. The argument that it has been eroded is invalid, as the uppermost stages have been recognized in the fault blocks, together with Triassic rocks (Mr J. B. Waterhouse, pers. comm.); under normal erosion, these upper stages should have been removed first. The hypothesis that the thickness has been diminished through faulting presumes that the Otago Schist and Fiordland-Stewart Island complexes have remained stationary. If this were so, these complexes would have strongly influenced the lithology in this bottle neck. No evidence of such a change has been found; on the contrary, a similarity between Southland and Nelson stratigraphies for the upper Paleozoic is constantly emphasized. Such close correlation of these stratigraphical columns would hardly be possible if the Fiordland and Otago complexes had always been in their present positions. The only satisfactory postulate would seem to be that the distance between the Otago Schists and the Fiordland Complex had, at an earlier stage, been greater than at present; either the Fiordland Complex was farther south or the Otago Schists farther north.

The concurrence of a large Bouguer anomaly with the Otago Schists (Fig. 6) would be accounted for by postulating a greater thickness of light crustal material underneath the Otago Schists. A massive thickness of strata, which could not be accommodated in the Eglinton-Hollyford valley, situated underneath the Otago Schists, would account for the anomaly and provide an explanation for the Eglington-Hollyford valley problem. This hypothesis would, however, entail the acceptance of an overthrust in the western and southern Otago districts. The presence of an overthrust on the southern side of the Otago Schists near Balclutha (Macpherson, 1946; Wood, 1956) supports such an assumption, particularly where this overthrust merges into a belt of ultramafics farther west. The position of a belt of peridotites, dunites, gabbros, and serpentinites along the front of the Otago Schists is very similar to that in other alpinotype structures (Jacob, 1954). De Roever (1957) has suggested that these ultramafics were placed in their present position as fragments of the peridotite crust, and not in a liquid state. The mechanics of such an implacement are not quite understood, but certainly require a large thrusting movement. The presence of this belt of ultramafics along the front of the presumed nappe supports such a hypothesis.

The author is of the opinion that these ultramafics were probably in a liquid state in an earlier stage of the overthrust and then solidified, solidification taking place before further overthrusting. Transportation, as presumed by De Roever (1957), probably occurred in later stages of the overthrusting, hence the presence of thrust planes as observed by Hutton (1937) on top and bottom of the ultramafic bodies.

The general swing in strike to the south-south-east in the Southland Synclinorium in south-east Otago and southern Southland (I. G. Speden, in press) also supports such an assumption, as it could well represent a return to the original strike of the geanticline.

FAULTING IN THE MARLBOROUGH PROVINCE AND ON THE EAST COAST OF THE NORTH ISLAND

As faulting in Marlborough is intimately related to that farther north on the east coast of the North Island, both regions are discussed together.

Piercement Structures

The development of small-scale piercement structures, associated with transcurrent fault zones on the east coast of the North Island, has been described by the author (Kingma, 1958c); it seems that these transcurrent fault zones have been active during most of Tertiary time, at least from as early as the beginning of the Oligocene. In these piercement structures, Cretaceous rocks are usually revealed. The fact that the Cretaceous sequence encountered in these piercements is the same as in the district surrounding the piercement structures is sufficient evidence that no exceptionally great horizontal movements have taken place.

The piercement structures and the consequent local unconformities thus came into existence locally and not all along the fault zone. If, for instance, Waitakian sediments are found to be resting on Haumurian rocks in the piercement, this is ample evidence that in immediately pre-Waitakian time, Haumurian rocks were being exposed by transcurrent movement along the fault. The clear-cut fault traces along the east coast transcurrent fault zones, and displacement of terraces and spurs along the Ruahine-Rimutaka horst, clearly indicate that movement is still in progress. The faults bordering the Ruahine-Rimutaka horst on the eastern side may be traced far to the north, but transcurrent movement has not satisfactorily been established there. All the other fault zones farther east terminate either in or south of the Hawke Bay.

Small piercement structures may occur in a transcurrent fault zone, while large-scale piercement structures may be found between two transcurrent fault zones. These large-scale piercements are, in the author's opinion, a very common feature in the geology of New Zealand They always result from shortage of space, and the wedge-shaped Ruahine-Rimutaka horst is one such piercement structure. The Marlborough Mountains (Seaward Kaikouras, Inland Kaikouras, Spenser

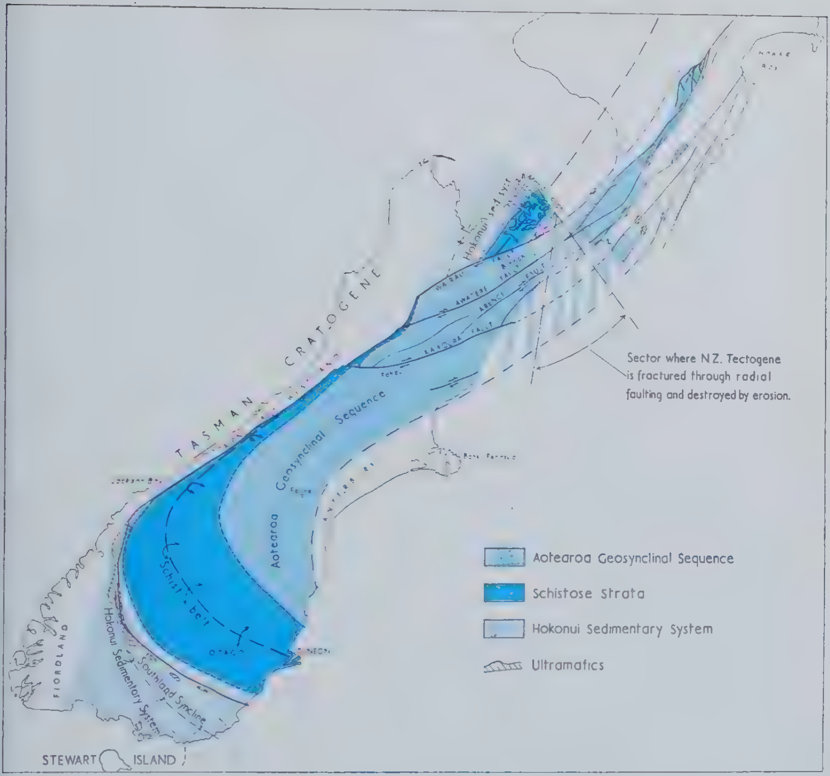
Mountains) are all separated by powerful fault zones which, as Wellman (1956) has shown, are all dextral transcurrent. Kingma (1957c) and Lensen (1958b) have presented evidence for joining the Marlborough transcurrent faults with those in the east coast of the North Island, and in Marlborough also it seems to be evident that, to maintain dextral transcurrent movement on all faults, the more easterly situated blocks have to move faster than the neighbouring westerly blocks.

The Aorangi Range (Fig. 1), which is a large-scale piercement structure, has suffered extensive internal deformation (pers. observ.). The greywackes which strike north-east with varying dips have been stretched parallel to the strike, causing extensive boudinage (Rast, 1956) of the sandstone bands and sliding in the finer grained rocks.

In the North Island, movement started early in the Tertiary as mild, almost insignificant displacements; these continued into Marlborough and piercement structures and small local unconformities are to be expected in Marlborough geology. The southern extension of the Ruahine-Rimutaka horst in Marlborough culminates in the Spenser Mountains, which are regarded as being similar to the horst. The Dannevirke-Masterton sector and its Marlborough counterpart, the Kaikoura sector, constitute another block moving between two transcurrent fault zones (Figs 1, 3). Where the zones change direction, as in Cook Strait, the block in between the zones will not, as it were, follow the guide lines (Figs 3, 9); it will "derail", and either override the guide lines or thrust itself into the air. The latter phenomenon is seen in Marlborough, where the Seaward and Inland Kaikoura Ranges point into the air as knife-sharp ridges, with the Tertiary cover still preserved in patches on the western sides. The sharp-crested Aorangi Range in the southern Wairarapa is a similar feature. The southward movement of these blocks during the Kaikoura Diastrophism was so great that on the northern side of Aorangi Range (Fig. 7) a deep hole was left behind which was filled during the Nukumaruan sedimentation. A similar hole was described by the author from Central Hawke's Bay (Kingma, 1958a), where, during Tongaporutuan time, very rapid sedimentation took place in small basins. The Taruarau Saddle (Fig. 5) and the Cook Strait rupture are similar features that resulted from southward movement of the East Coast-Marlborough segments.

Marlborough Stratigraphy

A wide-scale investigation of the Marlborough Province has not yet been undertaken, and a coherent geological history of this province is consequently not available. The information so far collected clearly points to the fact that the Marlborough district is situated in the same depositional belt as Canterbury and the east coast of the North Island, earlier in this account called the Eastern Basin. From work by Macpherson (1952), Wellman (1955a), and Suggate (1958), the correlation is evident. The transcurrent fault zones and possible associated piercement structures, mentioned earlier in this chapter, must have affected the sedimentation in Marlborough, just as they have influenced the



Sector where N.Z. Tectogene is fractured through radial faulting and destroyed by erosion.

FIG. 9.—Outline of the Alpine Schist Arc and its relation to the Marlborough-East Coast, North Island, segments.

sedimentation farther north in the North Island, but their effect was not great enough to cause serious discontinuities in facies. The main break-up of this area occurred during the Kaikoura Diastrophism, when the sharp Kaikoura ridges emerged and large quantities of conglomerate were produced, from which Marwick (1934) collected fossils showing clearly a Waitotaran age. The Great Marlborough or Bourne Conglomerate was linked by Marwick with the widespread marine transgression that invaded the eastern and western parts of the North Island in Plio-Pleistocene time.

A recent study (Suggate, 1958) of the geology of the Clarence Valley from Gore Stream to Bluff Hill reveals a stratigraphy that can easily be correlated with its counterpart in time on the east coast of he North Island and in Canterbury. An angular unconformity between he upper Jurassic (Heterian) and the Turonian (Motuan-Ngaterian) vas observed, and the area is transitional between North Marlborough nd North Canterbury, where quartzose coal measures, underlying narine Piripauan and Haumurian, rest with obvious unconformity on

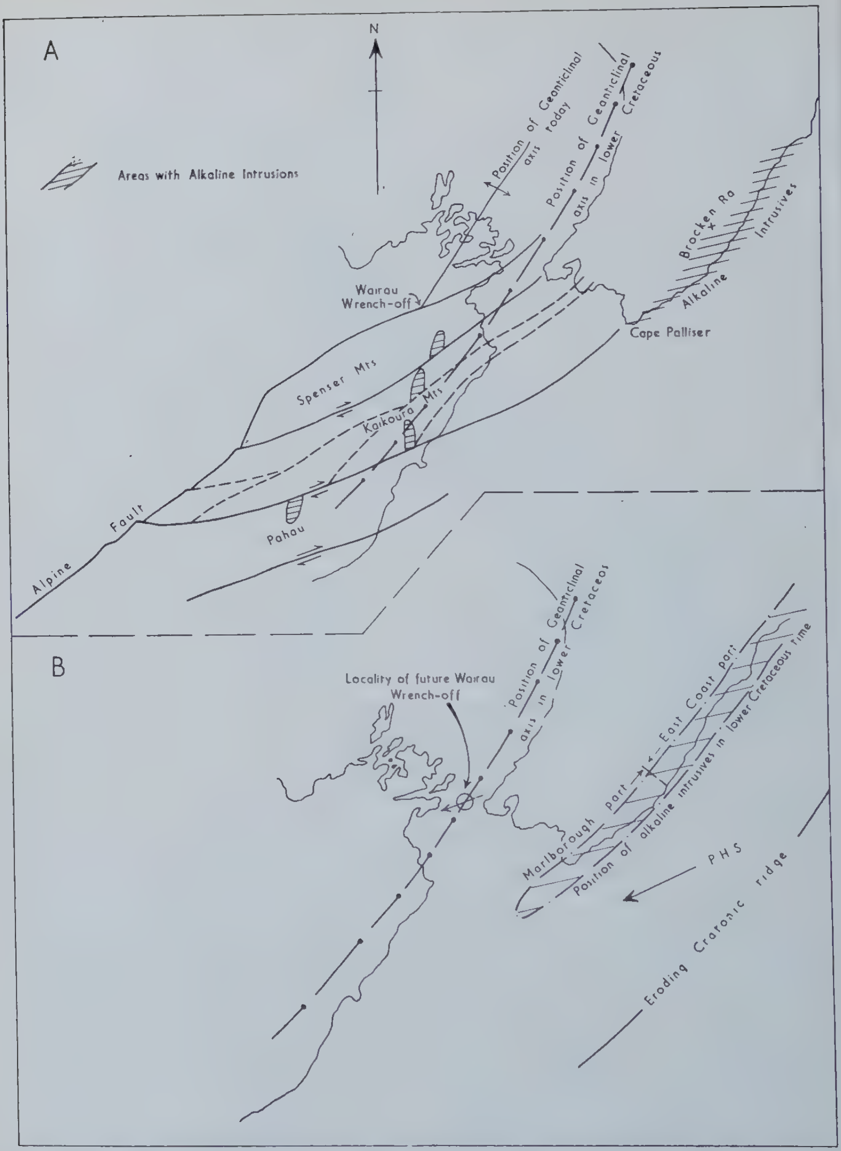

A
N
Areas with Alkaline Intrusions
Position of Geanticlinal axis today
Position of Geanticlinal axis in lower Cretaceous
Brocken Ra
Alkaline Intrusives
Cape Palliser
Wairau Wrench-off
Spenser Mts
Kaikoura Mts
Pahau
Alpine Fault
B
Locality of future Wairau Wrench-off
Position of Geanticlinal axis in lower Cretaceos
East Coast part
Marlborough part
Position of alkaline intrusives in lower Cretaceous time
PHS
Eroding Cratonic ridge

Fig. 10.—A: Position of the Alarm Intrusives in Quaternary time (Marlborough-East Coast). B: Original position of Alarm Intrusives during lower Cretaceous.

the Aotearoa geosynclinal sequence. The Marlborough area was brought above sea-level, folded, and eroded before the Turonian sedimentation and igneous activity commenced, while farther north in the North Island marine rocks were laid down, and farther south quiet terrestrial conditions prevailed. The orogeny is therefore an uppermost Jurassic to lower Cretaceous feature and younger than that in South Canterbury (Gage, 1957). The Turonian basalts, and the probably older alkaline intrusives which played such a prominent role during the deposition of sediments in Marlborough, are of utmost importance for the reconstruction of the tectonic movements in Marlborough. As long ago as 1890, MacKay pointed out that the basaltic bodies are always cut off by faults and noted the similar lithologies in the separate igneous bodies. The faults mentioned by him are the great Marlborough transcurrent faults. That the basalts were formed in their present location is highly unlikely, as the faults have been transcurrent for a long time and the basaltic bodies must have been affected by the movement along them. Dr H. J. Harrington (pers. comm.) suggested that, if the bodies are slid back along these faults, they would make one large united body in the northern part of Marlborough, perhaps in the Cook Strait area, or even farther north.

Wellman and Brodie (1954), investigating an area at Cape Palliser, discovered dykes of fine-grained, holocrystalline, porphyritic rock that has provisionally been classed by Reed as hornblende camptonite (alkaline lamprophyre), the plagioclase being more acidic than usual (P. 12873) (Reed, *in* Wellman and Brodie, 1954).

Hutton (1943) described teschenites and dolerites (P. 3230–1, 7601–2, 7608, 7616) from the Brocken Range north of Cape Palliser (Fig. 10). These rocks are probably comagmatic with the Cape Palliser camptonite and belong to one alkaline suite.

Mason (1958) described intrusive rocks from the Kaikoura Mountains and emphasized the mutual relationship between these intrusives and those investigated by him in the Pahau-Mandamus area in northern Canterbury. In the Seaward Kaikouras, the rocks are mainly gabbros, which generally consist of varying proportions of plagioclase, augite, and brown hornblende, commonly with some olivine, and usually with some accessory biotite (less than 5%). The plagioclase is sometimes practically absent, giving an ultrabasic variant consisting of approximately equal amounts of augite, olivine, and brown hornblende; the olivine is commonly partly or wholly replaced by chlorite and serpentine. Some pebbles of lamprophyre were collected, which proved to be camptonites containing abundant brown hornblende, generally some augite, and some olivine altered to serpentine and chlorite.

Similar intrusions of alkaline rock, carrying biotite and alkali feldspar in the gabbros, and alkaline augites and hornblendes in the syenites and trachitic rocks have been described by Mason (1951) from the Pahau-Mandamus area, and it has been found that these rocks are closely related to the Marlborough intrusives. Mason's explanation that these intrusives were originally closer together and drifted apart due to transcurrent movement along the main Marlborough faults is consistent with the theory expounded in this paper. Mason did not pursue the matter further, but some conclusive evidence can be derived from these intrusives. In the light of Mason's hypothesis, the intrusives may be considered to have been emplaced originally farther to the north-east. If the segments are moved back into their postulated positions during lower Cretaceous time, this belt of alkaline intrusives forms an extension of the alkaline intrusives of Cape Palliser and the Brocken Range (Fig. 10), in fact, of the whole sequence of lower Cretaceous intrusives of Hawke's Bay and the Wairarapa. The similar composition of these intrusives cannot be accidental, and it appears most appropriate to regard them as members of one and the same suite. As major outcrops of these igneous rocks may be found on Mount Alarm (9,400 ft) in the Inland Kaikouras, it is here proposed to call the intrusives the "Alarm Intrusives". The age of these intrusives is in all cases lowest Cretaceous, perhaps Jurassic (Mason, 1958); they should form convenient markers for areas with doubtful lower Cretaceous stratigraphy.

From the middle and upper Cretaceous of the east coast, igneous pebbles have been reported (Ongley, 1935; Lillie, 1953) and have been collected by the author. These pebbles have not been analysed, but they could very well have been derived from the Alarm Intrusives. The Pahau and Blairich segments are geographically the extreme of this suite in the South Island; the Pahau segment drifted farthest southward, while the segments in the Seaward and Inland Kaikouras seem to have remained behind, probably through internal fracturing of the Kaikoura segment along the Clarence and Kekerangu faults (Fig. 10). The break-up of the body of alkaline intrusions into smaller segments probably started after the Eocene, as the author has observed that, in the east coast of the North Island, transcurrent faults become active at this time.

The Marlborough Faults can be traced south-westward until they cut the Alpine Fault. The manner of cutting is clearly shown on the latest geological map of New Zealand (N.Z. 1:2.000.000) (N.Z. Dep. sci. industr. Res., 1958), in Fig. 9, and in Wellman *et al.* (1952); it demonstrates clearly the dextral transcurrent nature of the movement along the faults. Major faults on the concave side of the Alpine Range are to be found over a large area. The most important are the Waikari, Waihemo, and Two Thumb faults. The Waikari Fault follows the Waikari River in North Canterbury and dextral transcurrent movement has been observed (Mr D. Hamilton, pers. comm.); this fault clearly belongs to the Marlborough fault swarm. The Waihemo Fault, in the Moeraki district of South Canterbury, has sinistral transcurrent movement, in contrast to the Marlborough Fault movements (Mr D. Hamilton, pers. comm.). In Central Canterbury, the Two Thumb Range is severely faulted on the eastern side; the fault continues northward and southward over large distances and possibly joints the Waihemo and Waikari fault zones (L. Oborn, pers. comm.).

The Alpine Fault

Although the Alpine Fault is not strictly part of Marlborough geology, it is for convenience discussed here with the Marlborough faults (Fig. 11). The manner in which these faults cut the Alpine Fault make a 300-mile displacement along this fault, as postulated by Wellman

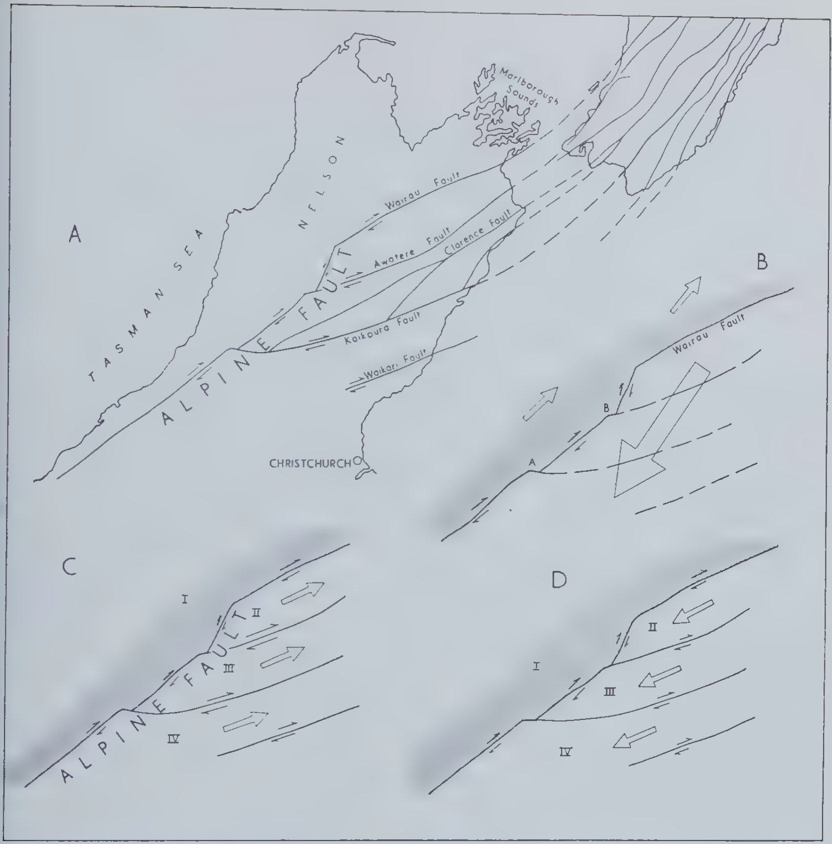


FIG. 11.—A: Outline of northern part of South Island showing Alpine and Marlborough faults; dextral transcurrent movement indicated by arrows along faults. B: Taking observed dextral transcurrent movement along Alpine Fault into account, block movement, as indicated by hollow arrows, would result in hypothetical rift forming along Wairau Fault and at points A and B. C: Taking observed dextral transcurrent movement along Marlborough faults into account, segments II, III, and IV would move away from segment I (Tasman Cratogene), causing hypothetical rift forming along Alpine Fault. D: Solution adopted in this paper; segments II, III, and IV move towards I (Tasman Cratogene), causing dextral transcurrent movement along Alpine Fault and on faults in between Marlborough blocks.

(1956), rather difficult to explain. The Kaikoura Fault is not the only one that reaches the Alpine Fault south of the Nelson district, for the Awatere Fault does the same, and would account for a proportion of the 300-mile displacement. This theory would have been feasible were the Alpine Fault straight. The fact that the Alpine Fault is cut by the Marlborough faults and divided into segments, and does not form a continuous line, is evidence that different types of movement took place on the Alpine and on the Marlborough fault systems. These considerations are, in the author's opinion, of sufficient significance to question the hypothesis that a 300-mile displacement has taken place along this fault.

The Alpine Fault undoubtedly exists, but detected lateral movement along it (Wellman, 1955b) is most probably the result of movement towards the west-south-west of the mountain chain. Field observations by Bowen (1954) relevant to this discussion show that glacial moraine has been overthrust by schist along the Alpine Fault. The fault was traced southward by Wellman and Willett (1942) as far as Jackson Bay and Lake McKerrow. The continuation of the fault south of this point has been inferred (Mr R. W. Willett, pers. comm.). The entire region between Lake McKerrow and the Pyke River is severely broken, and, apart from the continuation of the Alpine Fault through this area, there is in addition a dispersion into a large number of smaller faults radiating southward from the Pyke-Hollyford valley.

Another point to be taken into consideration is the fact that the Alpine Fault follows the Wairau Valley, and here a distinct change in direction of the fault is to be seen. Transcurrent movement along the fault, with the 300-mile displacement in mind, must cease as soon as the direction changes into the Wairau Valley, and the movement should be nearing normal. If this were so, an enormous rift should have developed along the Wairau Valley area. This is distinctly not so, and large-scale transcurrent movement is therefore unlikely to have taken place along the Alpine Fault. Tectonic displacement has been of a different nature, and movement of the Marlborough blocks in a south-west direction on to the Tasman Cratogene seems to have taken place. The position is outlined in Figs 11 A–D.

The Wairau Wrench-off

In Figs 10, 12, 15, 16, 17, and 19 the history of the Wairau Wrench-off is depicted. The large and spectacular displacement of the schist belt on the Wairau Fault must have had a long history; it reminds the author of the Judicaria Line in the Italian Dolomites (de Sitter, 1956; *see also* Rod, 1956). The displacement, together with the different attitudes of the schistosity planes in the Marlborough Sounds and in south-east Nelson (Reed, 1958), has led the author to the conclusion that the New Zealand Geanticline was wrenched off, probably in middle Cretaceous time. The lower Cretaceous, alkaline Alarm Intrusives have been extensively affected by this wrench-off, indicating that in lower Cretaceous time the wrenching had not yet started.

Prior to the wrenching-off, stretching of the schist belt was probably already at its maximum, so that with a shift in P.H.S. direction more towards east-west than in Jurassic time (Fig. 17), the geanticline was twisted off at the point of maximum strain. Once initiated, the displacement continued, accentuated by further shift in P.H.S. direction, the western component of the P.H.S. force gaining in magnitude at the expense of the southern component.

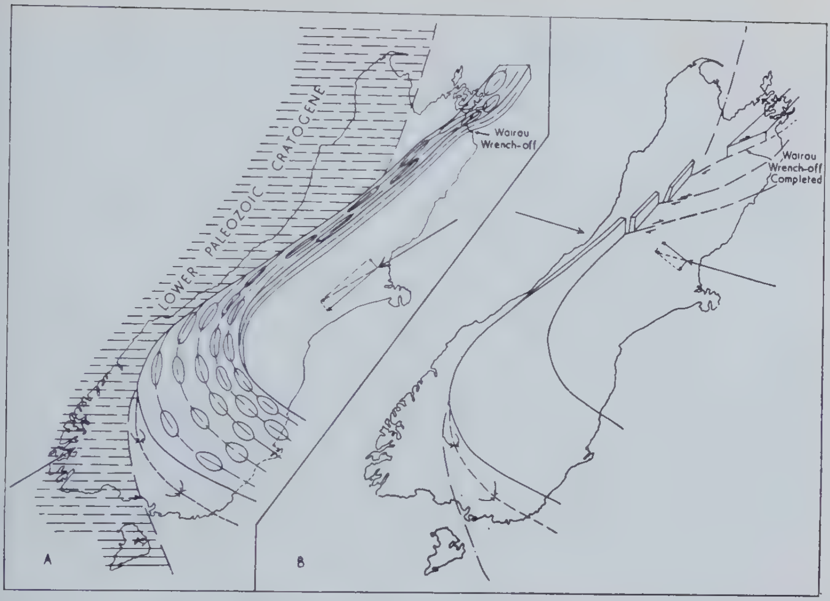


FIG. 12.—A: Outline of the New Zealand Tectogene in the latest stage of the Rangitata Orogeny. B: New Zealand Tectogene, broken up as a result of Kaikoura Diastrophism.

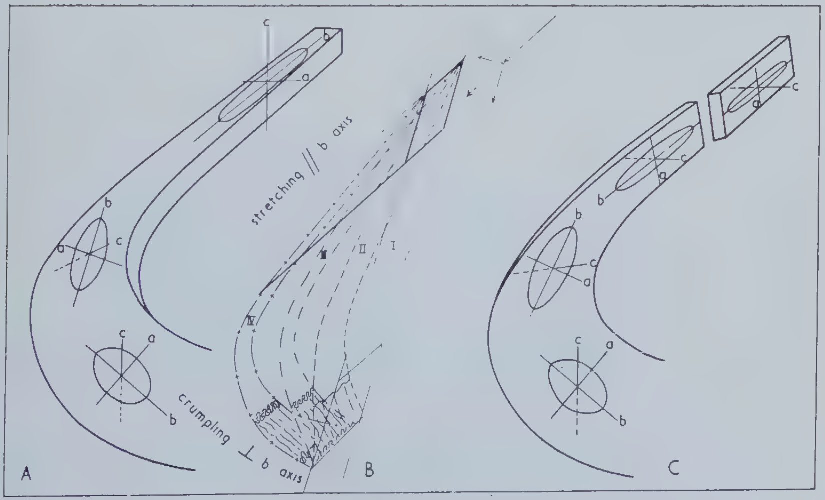


FIG. 13.—A: Position of strain ellipsoids in the New Zealand Tectogene, not including the Marlborough Sounds. B: Deformation of particles in the tectogene. Stretching in Southern Alps branch, crumpling (plus minor amount of stretching) in the Otago branch.

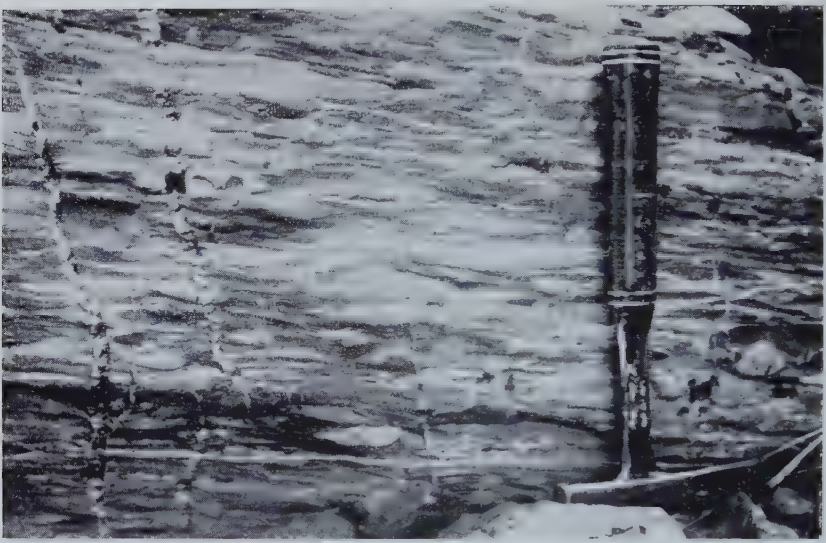
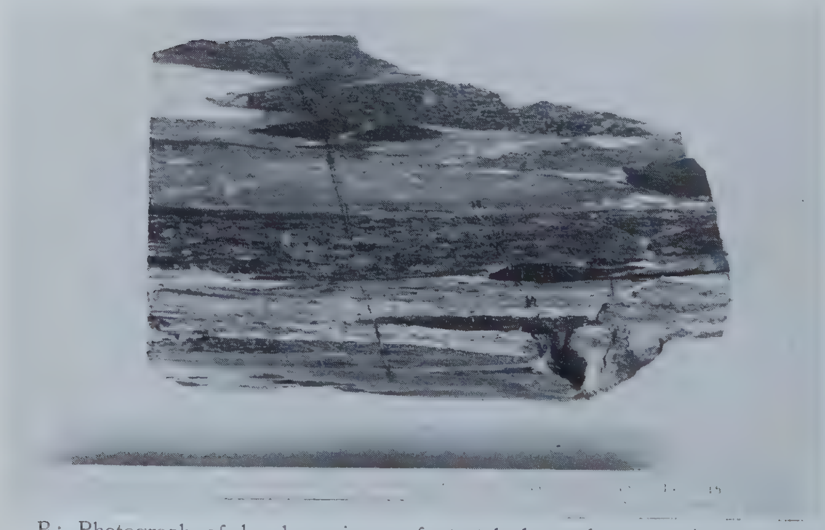


FIG. 14.—A: Photograph of block of stretched conglomerate, Stoney Creek, Two Thumb Range, Central Canterbury. (Black-and-white enlargement from Kodachrome.)

—*Photo*: *A. V. Weatherhead*



B: Photograph of hand specimen of stretched conglomerate (same locality as above).

—*Photo*: *S. N. Beatus*

Before the wrench-off was completed—i.e., before the geanticline snapped—stretching must have taken place in the geanticline farther north from the Marlborough Sounds area. This stretching and consequent collapse resulted in the drowning of the geanticline in that area (Figs 15, 19) and invasion of Tertiary strata in what later became the Wanganui Basin.

SCHISTOSITY IN THE ALPINE ARC

The Southern Alps-Otago Schist Arc is here for convenience called the Alpine Schist Arc. Distinctly different types of schistosity have been encountered in this arc, falling into two main categories. One category appears in the Southern Alps part of the arc and is a simple, usually vertical, schistosity, allowing slate-like separation along the schistosity planes; four zones may be recognized—non-foliated chlorite schist, foliated chlorite schist, biotite and garnet schist, and oligoclase schist—the metamorphism increasing westward (Hutton and Turner, 1936; Turner, 1936, 1940a). Metamorphism may be traced as far as south-east Nelson (Reed, 1958). Simple horizontal lineation, or lineation undulating about the horizontal, has been observed (Wellman *et al.*, 1952; Lillie *et al.*, 1957), due to simple schistosity without crumpling, probably resulting from severe stretching and slip. The schistosity planes are usually vertical and parallel to the bedding planes; beds of at least Triassic age have been affected.

The other main category has been observed in the Otago part of the Alpine Schist Arc and consists of intricately patterned microfolding. The schist as a whole has a very crumpled appearance; the microfolding has in places progressed so far that pencil-shaped rods have become separated. Apart from a certain amount of stretching and slip, the lineation is here mainly indicated by rodding (roling) (Wilson, 1953), and the schistosity planes are usually gently dipping. The Te Anau beds grade downward into the Otago schists without any sign of unconformity, and the schistosity has not affected young rocks such as those to be found in the Southern Alps.

It seems logical to accept that *plane of foliation* ≡ *schistosity* ≡ *ab plane of the strain ellipsoid*. Sander (1930), Billings (1954), Fairbairn (1949), Turner and Verhoogen (1951), Turner (1940b) limited the *b*-axis of the strain ellipsoid to parallelism with fold axes and often with lineation, the *a*-axis was perpendicular to the *b*-axis in plane of movement, and the *c*-axis perpendicular to the *ab* plane. Disagreement, however, still exists with regard to the lineations, the problem being with what axis of the strain ellipsoid the lineations should coincide.

Lineations parallel to the direction of movement have been observed by Reusch, 1887; Rüger, 1933; Kvale, 1945; Anderson, 1948; Billings, 1937; Mehnert, 1939. Lineations normal to the direction of movement have been observed by Sander (1930) and Schmidt (1924) (*see also* King, 1956). In the writer's opinion, both observations are correct, and transport either parallel or normal to the *b-axis* has to be taken into

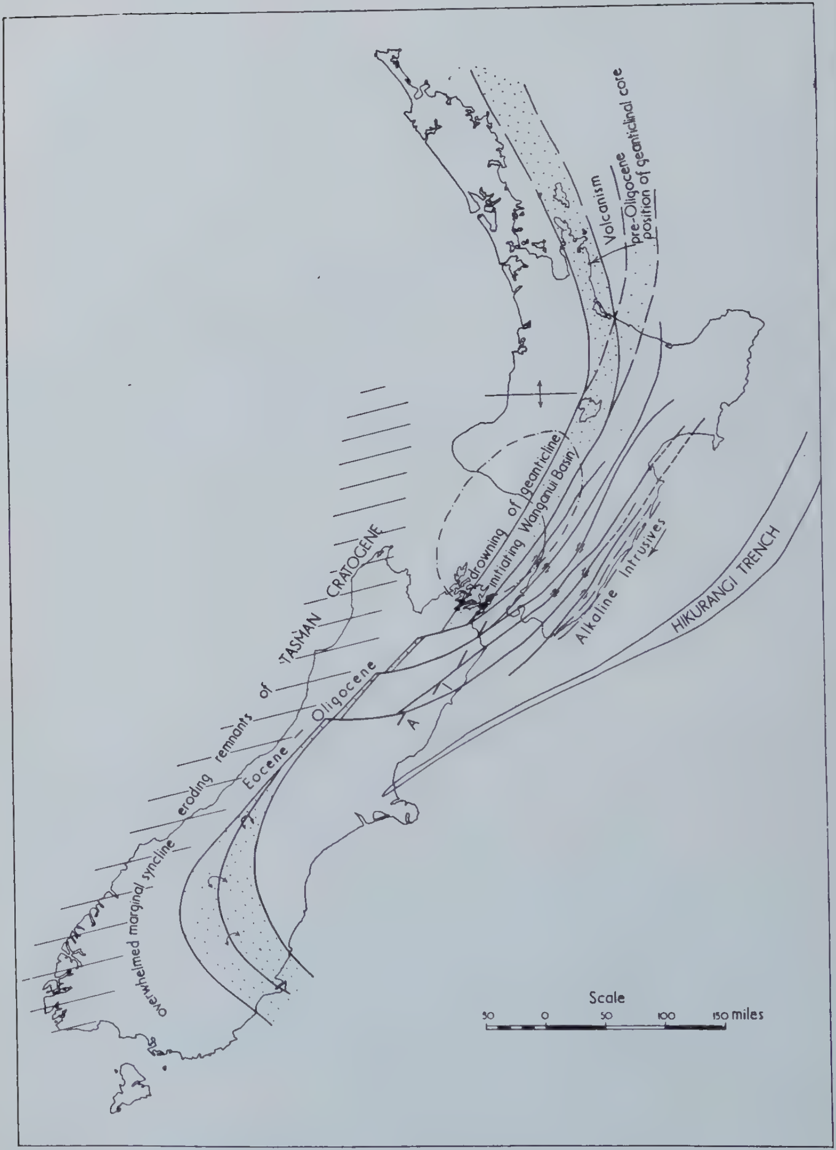

Volcanism
pre-Oligocene position of geanticlinal core
drowning of geanticline
initiating Wanganui Basin
eroding remnants of TASMAN CRATOGENE
Oligocene
Eocene
overwhelmed marginal syncline
Alkaline Intrusives
HIKURANGI TRENCH
A
Scale
50 0 50 100 150 miles

FIG. 15.—Outline of the New Zealand Geanticline (shown dotted) during the Oligocene. Fracturing of tectogene south of Nelson has taken place. Transcurrent fault zones active. North Island Geanticline submerging between Marlborough Sounds and National Park. Volcanism in North Island inferred from Poverty Bay Igneous conglomerates and tuffs.

account. The position to be ascribed to the strain ellipsoid depends entirely on the field occurrence of the lineations and the direction and time of movement for the particular area, preferably tested on additional data such as fold axes, slickensiding, nature of faulting, etc. In the Alpine Schist Arc, two different lineations have been observed. Both types of lineation in the schistose rocks seem to have attitudes commensurate with the direction of movement (P.H.S. in Figs 8 and 12 A), and do not present serious problems.

In the Southern Alps, the axes of folding have been clearly demonstrated (Wellman *et al.*, 1952; Lillie *et al.*, 1957; Waterhouse, 1955) they are more or less parallel to the general trend of the alpine chain, and axial planes are more or less vertical. The lithology of the unaltered rock is usually a banded system of alternating sandstones and muddy siltstones, generally termed greywackes, of the Aotearoa Geosynclinal Sequence. The schists are very probably altered greywackes in which the sandstones have been changed to quartzites and the muddy siltstones to schists. The greywackes grade into the chlorite schists.

The schistosity in the Otago part of the Alpine Schist Arc is, as already mentioned, of a very different type. Hutton (1900) drew attention to the general low dip of these rocks, although locally steep dips and intense crumpling have been noticed; he pointed out that their deformation was the result of regional thermal metamorphism. Daly (1917) called this type of metamorphism "load metamorphism". The author is of the opinion that schistosity only comes into existence through load metamorphism, when there is no lateral tectonic movement, as may be presumed for the schists in the graded sequences of the Waiuta series in Westland. The presence of corrugations, intense crumpling, and rod-like separations in the Otago Schists point to tectonic movement in a direction normal to these features, or, in other words, the lineation is caused by movement perpendicular to it. The observed lineation in Otago is usually horizontal, although locally the schistosity planes can dip steeply. This horizontal lineation is explained by Turner (1942) as micro-folding parallel to the *b*-axis of the strain ellipsoid, the *ab* plane being mainly horizontal.

From the foregoing considerations it is clear that tectonic transport in the two branches of the Alpine Arc is very different—e.g. parallel to the *b*-axis in the Southern Alps, and perpendicular to the *b*-axis in the Otago Schists (Fig 13 A, B, C).

A resolution of the Principal Horizontal Stress into components parallel to and at right-angles to the Alpine Chain shows the southward component to be greater than the westward, and consequently this greater southward component worked parallel to the Alps with only a minor force working in a westward direction. The result is a severe stretching and practically no crumpling by the westward force. Whatever crumpling may have developed was probably stretched beyond recognition.

In the Otago branch, just the opposite took place, the greater force working almost normal to the strike of the branch, resulting in microfolding and only a small amount of stretching. As the bending of the arc progressed, the position of the branches of the arc with regard to the P.H.S. direction altered continually, but stretching was always the main deformation in the Alps, while crumpling was predominant in the Otago branch.

In western Otago, where both systems of schistosity meet, a distinct change in lineation has been reported by Mr G. W. Grindley (pers. comm.). The lineation dips at about 45° to the south-west, in changing from horizontal in the northern Alps to horizontal in the eastern Otago Schists. In this area, Hutton (1940) found two distinct types of lineation, including a lineation perpendicular to corrugations, produced in the schists by a period of metamorphism antedating the movements that brought about the crumpling.

The situation is depicted in Figs 12 A and 13 A, where the probable attitudes of the strain ellipsoids are given. Following Kienow's (1953) analysis of internal rock deformation, the upper part of Fig. 13 B represents the deformation in the Southern Alps from a geanticlinal ridge initially more nearly north-south. The stretching obliterated any signs of previous deformation and crumpling is not likely to have played an important role. The deformation was probably rapid, with the stretching taking place under influence of a north-east–south-west principal horizontal stress. The deformation on the Otago part of the Alpine Schist Arc was a much slower process, and crumpling was a much more common feature than stretching. However, as this part of the Alpine Schist Arc attained a more responsive position with regard to the principal horizontal stress direction, stretching became an important feature, too. The Otago Schists probably went through a series of deformations with far less stretching than in the Southern Alps, but with successive phases of crumpling or lineation superimposed on one another. These different deformations and the S-planes recently revealed by the work of the University of Otago geology students (P. Robinson, 1958) are ascribed to subjection of the Otago area to stress, whilst rotating through an angle of 90° (Fig. 13 B, lower part).

The schists in the Marlborough Sounds area are at present being studied by Dr C. J. Vitaliano, of Indiana University, U.S.A. A general agreement with Otago exists, in that the schistosity planes, and consequently the *ab* planes of the strain ellipsoid, are horizontal, and in that the schists grade into strata of the Te Anau series.

Quartzose schists and slates are known from the Fairlie district (Fig. 9), where they have an attitude parallel to the Southern Alps and usually dip steeply. Crumpled schists, still clearly showing graded bedding, have also been found here and are obviously part of the quartzose schist-slate sequence. The age of these rocks is not known with certainty, but, as they appear well below graded beds with *Atomodesma,* their age is most likely to be pre-Permian. In this sequence a conglomerate bed has been found (Stony Creek, Two Thumb Range) that may prove

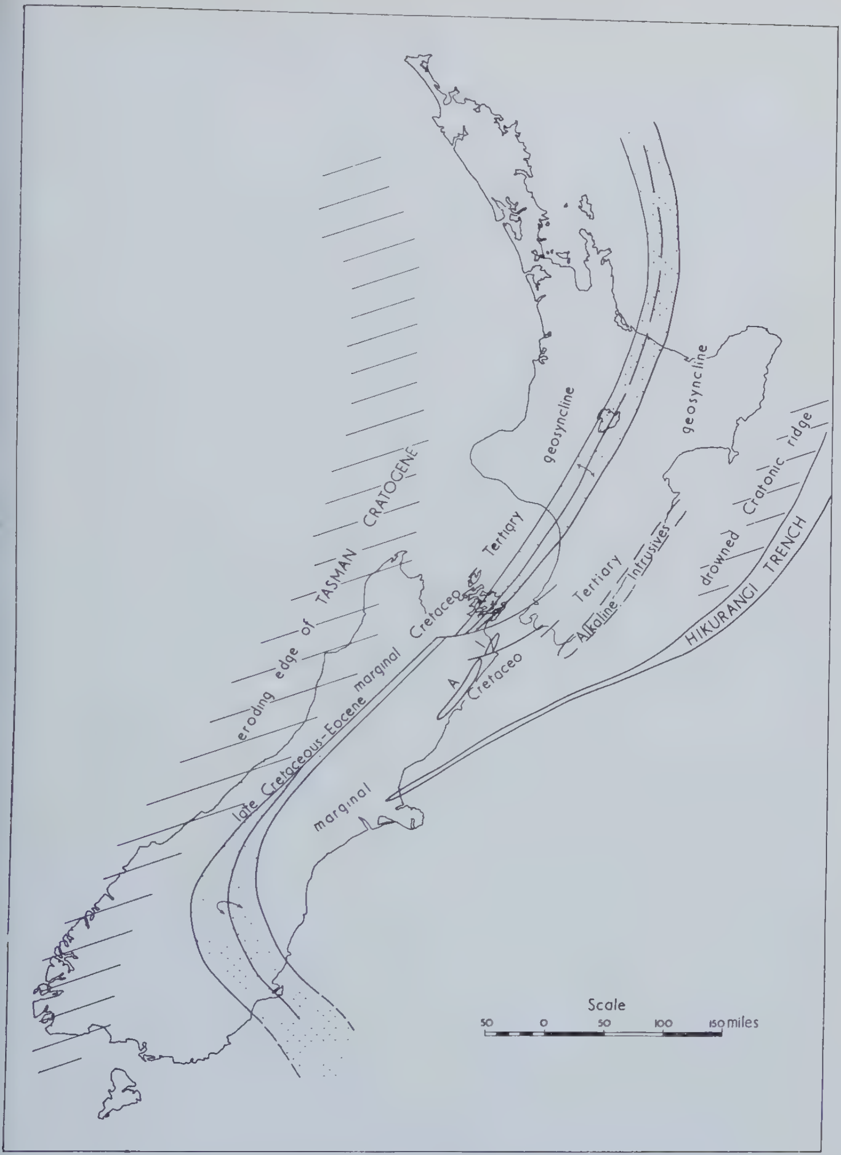


FIG. 16.—Outline of New Zealand Geanticline during the late Cretaceous and Eocene. Geanticline entirely emerged; Hikurangi Trench being formed, dragging eastern cratonic ridge down. Marginal Cretaceo-Tertiary basins developing.

to be of the utmost importance (*compare* Strand, 1945). The conglomerate is stretched, the components now having a rod-like appearance (Fig. 14 a, b). Sedimentary as well as igneous pebbles are incorporated

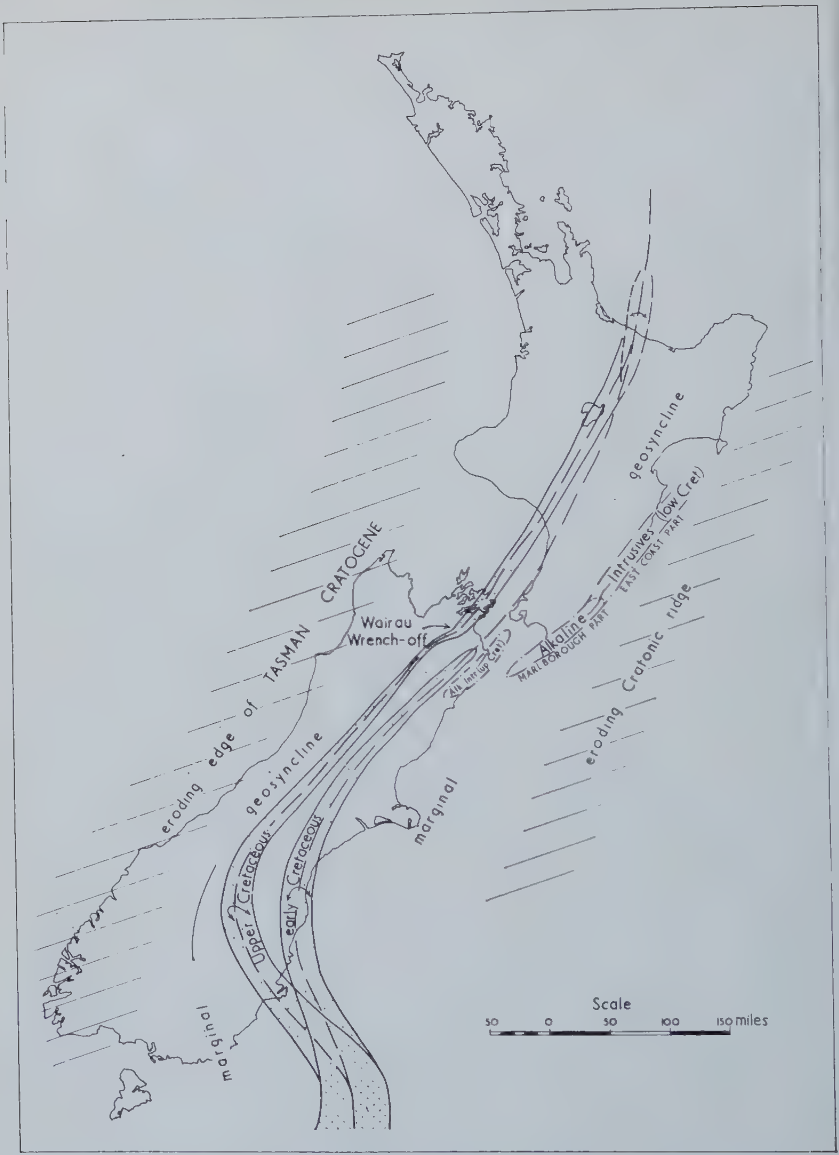


FIG. 17.—Outline of New Zealand Geanticline in the upper and early Cretaceous. Part of geanticline in Bay of Plenty area probably emergent during lower Cretaceous. Alkaline Intrusives emplaced.

in the conglomerate; the pebbles may be as long as 12 inches and half an inch to three-quarters of an inch in diameter. The rock was examined by Dr J. J. Reed, who reported as follows:

"A thin section of the Fairlie Conglomerate indicates a degree of internal reconstitution near the boundary of the chlorite Subzones 2 and 3 in the Classification of Hutton and Turner (1936), i.e., the greywacke pebbles are largely recrystallized, but a clastic texture is still shown by sparse relict large quartz and feldspar grains. The pebbles, whether derived from argillite, greywacke, or volcanic and tuffaceous material, retain their identity, as shown by different mineralogical composition, thus demonstrating insignificant chemical diffusion. The unbroken and sharp outline of the pebbles indicates that the amount of slip parallel to the schistosity planes is limited to the deformation of the pebbles, the schistose microtexture appearing to have formed primarily by crystalloblastic crystal growth perpendicular to simple compression. A linear structure is parallel to the elongation of the pebbles, i.e., in the direction of transport (*cf.* Anderson, 1948)."

The stretching in the Fairlie conglomerates is, in the author's opinion, substantial support for the theory of stretching in the Alpine Schist Arc. Stretched conglomerates have not so far been found in the Southern Alps strata, the facies being geosynclinal without coarse constituents. The author does not agree with Read's (1940) hypothesis that migmatization is the prime cause of regional metamorphism; he proposes the hypothesis that only a certain coincidence of load and movement, but primarily movement, causes regional schistose metamorphism, and that schistosity in the Alpine Schist Arc is mainly the result of deformation of the originally straight geanticlinal ridge under load. The deformation would, of course, be accompanied by mineralogical transformation.

OCEANOGRAPHIC DATA

Relief of the Tasman Sea Floor

Brodie (1952) discussed the sea-bottom morphology based on soundings shown on Admiralty Charts Nos. 788 and 1212, also on soundings taken by the *Dana* Expedition, 1919 (Greve, 1938). At the request of the New Zealand Oceanographic Committee, R.R.S *Discovery II* obtained a continuous echo-sounding record from Lat. 37° 02′ S, Long. 167° 28′ E, for 300 nautical miles to Lat. 39° 40′ S, Long. 172° 42′ E, off the western entrance to Cook Strait. From this data, contours were drawn and the dominant features of submarine relief established. The main features are the Lord Howe Rise and the Norfolk Ridge. The Lord Howe Rise continues the north-west trend of north-west Nelson, and the geological structure of this rise is probably very similar to north-west Nelson. The Norfolk Ridge lies well south-west of the line of the Northland Peninsula, and, unless it has been displaced, it is unlikely that this ridge is the continuation of the New Zealand Geanticline.

The Tonga-Hikurangi Trench

This trench can be followed from the Tonga Archipelago southward for more than 2,000 miles (3,500 km) and dies out against the South Island of New Zealand with a sweeping curve. It consists of three parts:

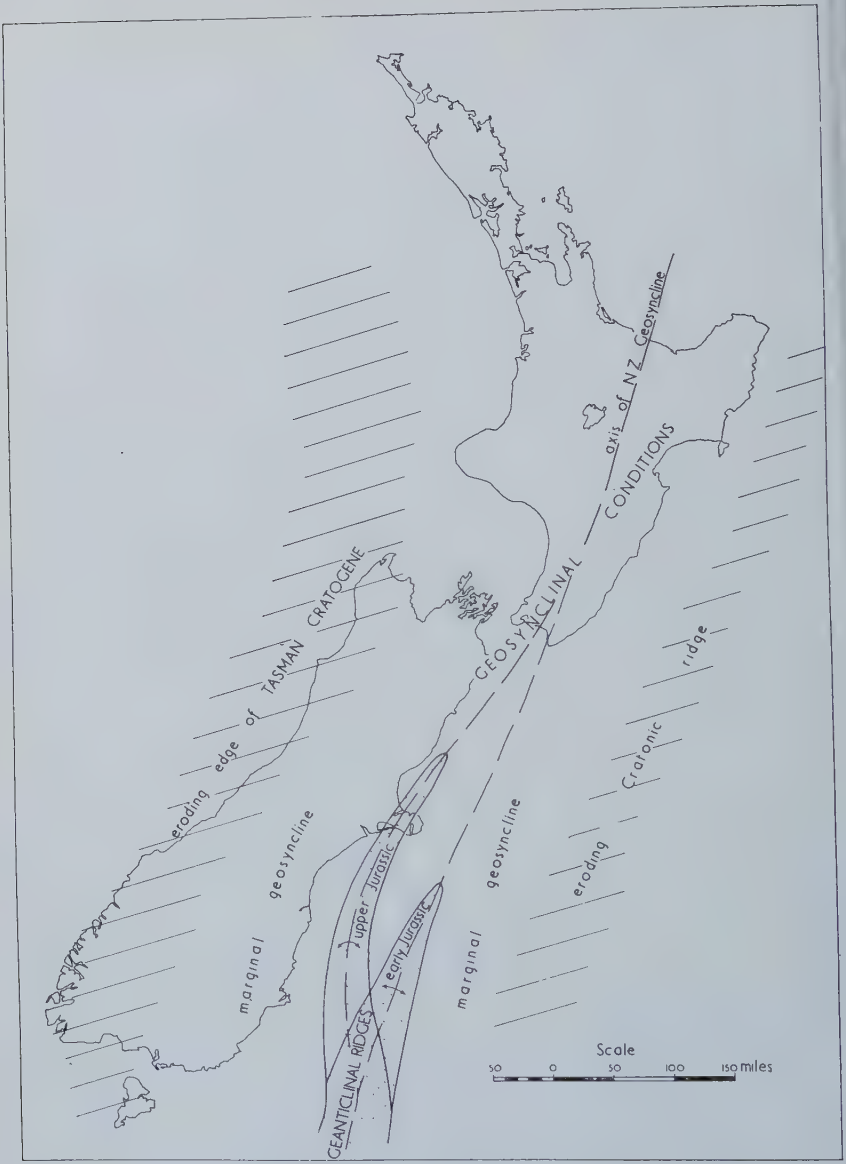


FIG. 18.—Outline of New Zealand Geanticline during the upper and early-Jurassic geosynclinal conditions still prevailing in North Island area. Tasman Cratogene extending much farther northward and probably westward.

the Tonga, Kermadec, and Hikurangi parts (Brodie and Hatherton 1958). Benioff (1949, 1954) has described this trench and its dee crustal structure as deduced from seismic data. The crustal buckling which, according to Vening Meinesz (1948), causes the negative anomalies over such trenches is accompanied by "orogenic" type faults; and the fact that deep-seated earthquakes occur in areas adjacent to these trenches, but always away from the downthrown side of the fault, is sufficient evidence that dislocation as visualized by Benioff exists. Oceanic faults mentioned by Benioff (1954, table 1) indicate that the dip of the fault plane of the Kermadec thrust is 58° west in the Tonga part and 64° west in the Kermadec part; the trench is clearly divided into three parts with different strikes. The significance of this dislocation is not fully understood, but is probably linked with the Kaikoura Diastrophism. The trench is probably a geologically young feature, as it is not filled with sediment to any appreciable extent; its western wall is sharply step-faulted (Mr J. W. Brodie, pers. comm.; and Dr G. B. Udinstev, pers. comm., U.S.S.R. *Witjaz* Research Vessel). The author's hypothesis as to the commencement of movement resulting in this trench is outlined in the last section of this paper. A gravimetric profile over the Hikurangi Trench has shown an absence of isostatic anomaly, and the trench appears not to influence the gravimetric field (Mr R. Dibble, pers. comm.). A full gravimetric survey of this trench is very desirable.

The Mernoo Bank-Chatham Island Rise

Fleming and Reed (1951) have described this feature in some detail. The Hikurangi Trench (Brodie and Hatherton, 1958) lies in between the Mernoo Bank-Chatham Island Rise and New Zealand; the trench dies out where this rise joins New Zealand—that is, at Christchurch.

The Chatham Islands, where an east-west lineation in the schist has been found (Mr A. Mutch, pers. comm.), are probably the remnants of an inner arc running parallel with the New Zealand Tectogene. This arc was probably destroyed with the development of the Hikurangi Trench, but, prior to its destruction, was probably continuous into the ridge which was situated off the east coast of the North Island. The Rangitata Orogeny, which affected the New Zealand Tectogene so much, most probably also affected the inner arc in the same sense, causing lineation in the schists to run in the same direction as in Otago. The Kaikoura Diastrophism probably caused the same type of movement in the Chatham Islands as in the Dunedin area, with rupture and accompanying volcanism.

SOUTH ISLAND VOLCANISM

Volcanism in the Marlborough district, mentioned previously, consisted of basaltic outpourings in the Cretaceous, and resulted in an extensive dyke system and in volcanic rocks associated with geosynclinal deposits. The South Island Volcanism considered here is situated on the concave side of the Alpine Schist Arc. Schofield (1951) described the volcanism in the South Island from the Upper Eocene to the Miocene; all eruptions were submarine and basaltic.

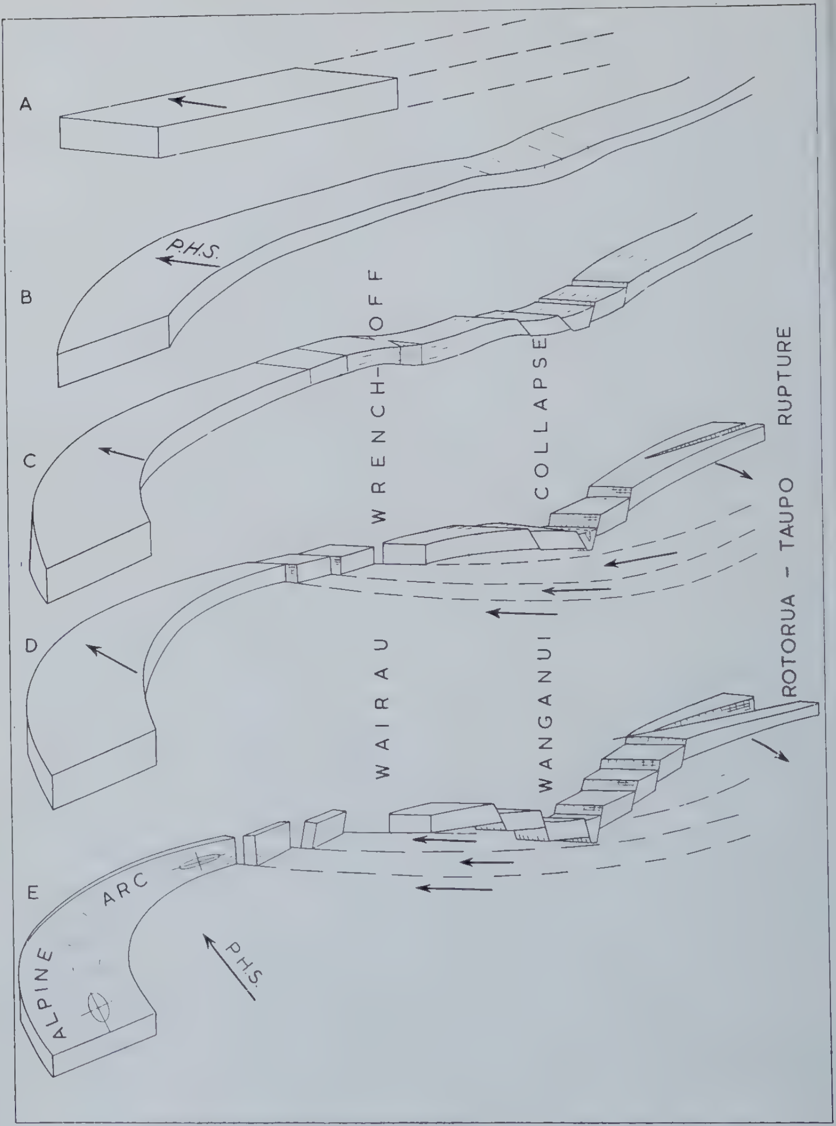


FIG. 19.—New Zealand Tectogene shown as a simple block. A–E various stages of deformation when block is subjected to a P.H.S. force from the northeast. A.B.C., internal deformation; D.E., external deformation.

A late Tertiary (Pliocene) period of volcanism can be recognized in the Dunedin district (Benson and Turner, 1945), where there were basaltic-trachitic outpourings. Banks Peninsula (Speight, 1944) has, apart from a mid-Tertiary volcanic episode producing rhyolites and pitchstones, a Plio-Pleistocene andesitic basaltic volcanism. The occurrence of this latter volcanism is, in the author's opinion, comparable with volcanism on the concave side of island arcs, as may be seen in the western Pacific and in Indonesia (Umbgrove, 1947).

THE RANGITATA OROGENY IN THE SOUTH ISLAND

As indicated earlier in this paper, the name Rangitata is proposed to replace the confusing name post-Hokonui. The Rangitata Orogeny, which was a mountain-building movement affecting deposits in the New Zealand Geosyncline, has been recognized in various places in the South Island. The general finding is that the Orogeny took place between the lower Jurassic and the upper Cretaceous, depending on the geographic position.

Wood (1956) placed the orogeny between the end of the Jurassic and the beginning of the Oligocene in the Gore Subdivision. This orogeny, although resulting in an angular unconformity, was apparently not severe. In Westland, Suggate (1957) accepts a severe orogeny in the earliest Cretaceous, but folding was still taking place in the late Cretaceous.

In North Otago, there is evidence of an orogeny, but the timing is uncertain. Gage (1957) postulates a vigorous orogeny after an accumulation of large amounts of sediment in the Carboniferous to Triassic, followed in the late Jurassic-lower Cretaceous by prolonged still-stand and erosion, to account for the development of the widespread Cretaceous peneplain before the close of the Cretaceous. In the uppermost Cretaceous, epeirogenetic movements allowed the flooding of the peneplain and consequent deposition of sediment. The timing of the Rangitata Orogeny is here uncertain, and to develop and run its course before the end of the Cretaceous would probably have commenced farther back in the Jurassic.

Mountain-building movement assigned to the Rangitata Orogeny in central Canterbury were completed before the middle Cretaceous (Suggate, 1958) and probably started in the upper Jurassic. In North Canterbury and southern Marlborough, the orogeny probably started rather later and had died down by the upper Cretaceous. In northern Marlborough, it was mainly an upper Cretaceous event, if it can be recognized at all (Macpherson, 1952; Wellman, 1955a). Although there were irregularities in deposition throughout the east coast of the South Island, a survey of the unconformities clearly shows that the rock underneath the unconformity gradually increases in age from northern Marlborough to Otago. The orogeny clearly started in the south, and, while the geanticline was emerging there, sedimentation continued in the north.

THE KAIKOURA DIASTROPHISM IN THE SOUTH ISLAND

The Kaikoura Diastrophism in Southland was hardly noticeable as a folding movement. Sedimentation in existing pre-Pleistocene basins continued in Westland, with huge quantities of sediment eroded from the emerging Alpine Schist Arc. A terrestrial sedimentation resulted in widespread terrace systems with terrace deposits in some places several hundreds of feet thick. Throughout the remainder of the South Island, the Kaikoura Diastrophism was contemporaneous and can be recognized as a strictly Plio-Pleistocene event. In places a mild pre-Pliocene folding movement has taken place, but these movements may have been connected with the dying stages of the Rangitata Orogeny, which lasted in the North Island until the beginning of the Miocene.

The main events comprising the Kaikoura Diastrophism were: the emergence of the Alpine Schist Arc and further rapid subsidence of the Canterbury Basin, the emergence of the Kaikoura Fault blocks, and the formation of Cook Strait. Simultaneously with the development of the Schist Arc and the Marlborough blocks, fault blocks developed on the inside of the Schist Arc, emerging along faults more or less parallel with the arc, and forming a set of west-dipping dip slopes. In these fault blocks, the pre-Permian stretched conglomerates are evident in the lowest parts of the east-facing escarpments. The attitude of these blocks is very similar to that of the Marlborough blocks, as they, too, have Tertiary strata preserved on their western, dip slopes. Sinistral transcurrent movement observed on the Waihemo fault zone (Mr D. Hamilton, pers. comm.) and to be expected on the overthrust south of Dunedin points to an east-west stress direction.

SUMMARY OF TECTONIC EVENTS

In the foregoing discussion, the main events and features in the tectonic and geological history of New Zealand have been outlined. These are undoubtedly related to one another and are the results of certain crustal movements that, once set in motion and by whatever process, continued to develop. The results of these movements are now to be seen in the geological formation of the country, and therefore, by logically tracing back New Zealand's geological history, it should be possible to fit them into their appropriate places in the sequence of events; so much evidence is available that it is felt that a reasonably well-founded history can be reconstructed.

The following are listed (not in chronological order) as the main events in New Zealand's history, and an attempt is made to weld these events into a co-ordinated account.

Development of the Wanganui Basin
Development of the Petane Trough
The Plio-Pleistocene sedimentation in New Zealand
The pre-Pleistocene sedimentation in New Zealand
The appearance of the Ruahine-Rimutaka horst
The appearance of the Aorangi horst
The presence of the Kaimanawa Ranges
The presence of transcurrent fault zones in the North Island
The presence of different earlier and later fold trends
The development of the Rotorua-Taupo volcanism
The volcanism in the east coast of the South Island

The appearance of the Marlborough Ranges
The presence of the Marlborough dextral transcurrent fault zones, and other fault zones, some showing sinistral movement
The presence of the Alpine Fault
The Cretaceo-Tertiary sedimentation in Canterbury and Westland
The presence of schistosity in the Alpine Arc
The presence of schistosity in the Marlborough Sounds
The appearance of gneisses and granite blocks in the lavas of National Park, and the Alarm Intrusives in Marlborough
The stretched conglomerates of Fairlie
The Paleozoic-Mesozoic sedimentation in Southland
The folding of the Southland Syncline
The presence of the Fiordland-Westland Cratogene
The presence of the Hikurangi Trench
The distribution of the Hokonui and Aotearoa Sediments
The zoning of the early-to-late Cretaceous unconformity along the east coast of the South Island
The position of the gravity anomalies

THE KAIKOURA DIASTROPHISM IN NEW ZEALAND

The Kaikoura Diastrophism, which took place contemporaneously throughout the New Zealand region, can be summarized as follows: The effects of the diastrophism were cataclysmic and superimposed on the dying Rangitata Orogeny. In the North Island, the diastrophism started at the end of he Miocene, when Kapitean sediments invaded the North Island Geanticline and filled the Taruarau Saddle, clearly showing that the deformation of the Wanganui-Petane depression had started. In this depression during lower Pliocene (Opoitian) time, the deformation continued, probably slightly accelerated, and a continuous belt of Opoitian sediments was laid down from Wanganui-Taranaki across the Taruarau Saddle into Hawke's Bay. During this deformation, the wedge-shaped Ruahine-Rimutaka horst was elevated and no sedimentation took place on the horst. In the following upper Pliocene (Waitotaran) time, movement continued along the same lines as during the preceding Opotitian, but in the uppermost Waitotaran the horst was submerged as far south as the Manawatu Gorge, allowing sedimentation to transgress the horst. This movement can be explained in the following way: During the formation of the Ruahine-Rimutaka horst, dextral transcurrent movement along the bordering faults took place and the en echelon segments of the horst moved southward. During the Opoitian, faulting in the Taruarau Saddle was already under way, and the Ruahine-Rimutaka horst became separated from the Kaimanawa-Kaweka Block. In the upper Waitotaran, when the horst submerged, this separation was completed and the horst probably settled down temporarily after its connection with the Kaimanawa-Kaweka Block was severed.

Like the Ruahine-Rimutaka horst, the Aorangi horst moved southward, as may be judged from the fact that it is situated in between transcurrent fault zones, and that a basin developed on its northern side (Fig. 7). In the Poverty Bay district, the north-east-trending Kaikoura deformation was superimposed on an earlier north-west-trending system, causing severe disruption of existing anticlines and synclines and the consequent development of piercement structures where anticline met anticline, and drop-ins where syncline met syncline.

The most spectacular result of the Kaikoura Diastrophism is to be seen in the central part of the North Island, where the island virtually split in two and enormous quantities of volcanic material were ejected. The accompanying faulting affected large areas on the east coast of the North Island. The general effect of the Kaikoura Diastrophism on the North Island was thus mainly accelerated movement on the transcurrent fault zones and folding with a north-east trend. Practically the whole Wellington province moved southward as a result of this accelerated movement, developing the zone of tension from Wanganui to Napier (Wanganui-Petane Trough). This zone of tension corresponds closely with the Rangitikei-Waiapu Anomaly (Fig. 6; Robertson and Reilly, 1958) and with the zone of tension postulated by Lensen (1958a). The cause of these drastic movements lies in a change in principal horizontal stress direction farther towards the east (Fig. 8), which did not, however, cause cessation of displacement on the transcurrent fault zones; that is to say, the angle between stress and fault zone nowhere approached 90°, and the remainder of the North Island behaved as a stationary block.

Apart from the emergence of the Alpine Schist Arc, the effect of the Kaikoura Diastrophism on the South Island is particularly noticeable in the Marlborough province, where the Kaikoura Ranges emerged. The Marlborough Ranges are all situated between transcurrent fault zones, and as such are the continuation of the east coast segments. Where these fault zones changed direction, the nature of the fault altered; the fault blocks "derailed" and were eroded into their present shape. Where they changed direction, radial faults must have developed, and the formation of Cook Strait probably took place along these radial faults (Fig. 9). Some escarpments which could be explained as being the result of such faults have actually been discovered (Dr H. Pantin, pers. comm.). The Kaikoura Diastrophism in the South Island induced an intensification of already existing features and consequently the Marlborough segments shifted farther south-west, cutting the Alpine Fault into a number of sections. The largest displacement has taken place along the Wairau Fault, but not all this movement is to be attributed to the Kaikoura Diastrophism, as will be shown in the following discussion on the Rangitata Orogeny.

The Rangitata Orogeny in New Zealand

When the effect of the Kaikoura Diastrophism is discounted, and the deformation caused by this revolution subtracted, it appears that most of New Zealand today would be below sea-level. The segments in the East Coast-Marlborough area would be moved back to the north, as would the Kaikoura Ranges, partly filling in Cook Strait. The displacement along the Marlborough faults would not have been so great, and the New Zealand Geanticline would run across the Wanganui

Bight towards National Park. This, then, would be the arrangement in lower Tertiary time. It has been seen that during this time the transcurrent fault-zones in the Marlborough-East Coast area were already active, and rupture of the New Zealand Geanticline was already in progress where the Marlborough faults now cut the Alpine Fault. There is little evidence that the transcurrent fault zones were active in pre-Oligocene time, and during this time the geanticline was probably still intact. From a survey of the schists in the Alpine Schist Arc, it has been concluded that stretching took place in the Alpine arm of the arc; in late Cretaceous to Eocene time this stretching must have reached its maximum; probably during this time the last schists were formed. The different attitudes of the *ab* planes in the schists of Marlborough Sounds and in those of the northern part of the Alpine Schist Arc were brought about during this stretching. At the same time the Alpine Schist Arc was torn off the northern part of New Zealand Geanticline (Wairau Wrench-off), causing a higher regional metamorphism in the Alpine schists (Fig. 12A) through greater internal movement. The arc ruptured during subsequent movement and the displacement was emphasized by the Kaikoura Diastrophism. The rupture caused stretching farther north in the geanticline, hence the development of the Wanganui Collapse (Fig. 19). This collapsed area is now occupied by the Wanganui Trough. The formation of this collapse was caused solely by the movement of the Alpine Schist Arc. The Wanganui Collapse was thus purely a mechanical development and not the result of local geochemical action in the substratum. The tectonics involved in the development of the Wanganui Collapse strongly resemble those found experimentally by H. Cloos (1929, 1930; *see also* de Sitter, 1956). The extensive boudinage and slip observed in the Aorangi Range (Fig. 1) would probably have developed in a higher degree of schistosity had pressure and depth of burial been greater during the deforming movements.

Farther back in the history of the New Zealand Geanticline, the elimination of the stretching is required and, consequently, straightening of the geanticline. As the geanticline is straightened, the Alpine Arc will reach less far west, and the marginal syncline, of which now only parts are visible in Southland and Nelson, will appear. If this hypothesis is accepted, the Alpine Schist Arc is nothing but a great nappe. The force causing this overfold is probably a rather shallow, epidermal thrust, generally directed from the north-east towards the south-west. The recent finding that the Hikurangi Trench has no isostatic anomaly (Mr R. R. Dibble, pers. comm.), and consequently that no great deformation of the Mohorovicic Discontinuity is to be expected, may be explained by this epidermal thrust. The disappearance of the eastern cratonic ridge in the upper Cretaceous (Fig. 16) is probably to be connected with this thrust, as the initial stage of the development of the Hikurangi Trench. The effects of the Rangitata Orogeny will be more and more nullified farther back in the history until the original New Zealand Geosyncline is re-established. In this geosyncline, sedimentation on the western side—the Hokonui Sedimentary System—was predominantly volcanic; the sedimentation on the eastern side—the Aotearoa Geosynclinal Sequence—was almost non-volcanic, without limestones and predominantly with graded beds; these Aotearoa sediments now predominate in the New Zealand mountain ranges.

The history of the New Zealand Geanticline is depicted in Figs 15 to 18. In Fig. 19, the development of the New Zealand Tectogene is shown as the deformation of a simple block. The New Zealand Geanticline is for convenience trimmed into a simple cube and subjected to a P.H.S. force from the north-east. The Alpine Schist Arc, the Wairau Wrench-off, the Wanganui Collapse, and the Rotorua-Taupo Rupture are shown as a natural result once the P.H.S. force was initiated.

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PALEONTOLOGICAL CASTING AND MOULDING TECHNIQUES

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Summary

Methods of making rubber casts from natural moulds, and plaster replicas of fossils using plastic moulding compounds are described.

INTRODUCTION

The ability to produce exact moulds, and so reproduce perfect replicas of important fossils, has long been recognized as a worth-while and important technique in paleontological studies. Plaster casting makes available, to research workers and institutes all over the world, exact copies of unique specimens which under normal circumstances would not be available for loan. For exchange purposes, plaster replicas of important fossils are generally more acceptable than inferior duplicates, and casts of unusual and valuable specimens can be sent to specialists for comparison and comment without risk to the original. For museum displays, carefully painted replicas can be used when original specimens are not available or are too heavy for attachment to vertical panels.

During recent years, new moulding materials have become available which now make plaster casting a simple but exact process. The object of this article is to describe the methods and materials used by the Paleontological Section of the New Zealand Geological Survey in producing moulds and plaster replicas, since experience has shown that these are easier and quicker than older methods.

IMPRESSION COMPOUNDS

Because many fossils occur as natural moulds, impressions must be taken from them, to allow them to be studied as positives. For many years, the only impression compounds available for this work were various types of waxes and plasticine. The use of these for laboratory preparations is strictly limited. Waxes, when set, are very brittle, and the degree of penetration is limited by the necessity of easy extraction. Neither wax nor plasticine can give a satisfactory impression from a mould with deep cavities and re-entrant angles.

Waxes and plasticine have now been superseded by various types of rubber latex compounds. At times, however, the former may still be useful in the field for taking impressions from moulds exposed in friable, broken, and extremely hard rock before attempting the difficult removal of the fossil. Liquid rubber compounds are now widely used throughout the world to make impressions and for preparing moulds for casting fossils. Casts can also be made of delicate zoological specimens by the use of latex moulds (Pape, 1955).

N.Z. J. Geol. Geophys. 2 : 56-65.

In the New Zealand Geological Survey, the use of rubber latex is almost entirely restricted to the taking of positives from natural moulds of fossils. The method, using latex to make moulds for plaster casting, as described by Rau (1955), is only used occasionally, as another method, described below, has proved easier.

The latex used is supplied by Rubber Distributors Ltd., Wellington, in pint, quart, gallon, and four-gallon tins, under the name "Compounded Moulding Latex", available only for strictly non-commercial purposes. It is a rubber base dissolved in a mixture of ammonia and water, with a colouring pigment added, giving a pink appearance. (Various other colours can be obtained, but pink seems to be the most suitable colour.) On setting, the colour changes to a light red. If a tin is left open, the ammonia soon evaporates, leaving a thickened solution, but this can be returned to its former condition by the addition of ammonia and water. The latex appears to have a corrosive action on the tins, so large stocks should not be accumulated.

Before latex is applied, the natural mould of a fossil should be cleaned of any matrix or dust that may be covering fine sculptural details. This accomplished, a gentle brushing with water makes the surface ready for the latex. For ease of penetration and flowage, the latex is best applied to a wet surface, since wherever the water has penetrated, the latex will follow. Weak detergent improves penetration into cavities. Normally, a single layer of latex at room temperature takes from three-quarters of an hour to an hour and a half to dry, but the drying process can be speeded up considerably by using an electric hair drier. With the warm draught directed on the latex, solidification can be obtained within fifteen minutes. Care should be taken to avoid applying the latex too thickly, as pools will result in some areas and a skin will form due to solidification on the surface, which prevents further drying from taking place.

A small brush is often useful for applying both the wetting agent and the latex to specimens with complicated surface sculpture. However, after use with latex, the brush should immediately be washed in ammonia solution to prevent clogging of the hairs. Normally a metal or wooden spatula is satisfactory. Special care is needed to avoid bubbles and to ensure maximum penetration when applying the first layers, as these will carry the impression of the fossil, the succeeding layers only adding strength.

Some research workers have found a simple vacuum pump useful for inducing the latex into fine cavities (Gill, *et al.*, 1956), but with most of the work undertaken locally, careful wetting of the surface prior to applying the first coat of latex has produced the most detailed positives (Figs 1 and 2). After the first layer has completely dried, the second can be applied straight from the tin. Numerous coats of latex will produce a firm and flexible positive. The addition of cotton wool or loosely woven cloth between the layers will strengthen the cast. The light red colour is quite satisfactory for most purposes, but colours can be modified by the addition of water-soluble dyes, paints, or inks to the latex. A little india ink produces a black cast very suitable for photography after "fuming" with ammonium chloride. Two-tone positives, where the fossil is a different colour from the background matrix, although not necessary for scientific purposes, can be useful for educational and display purposes (Fig. 1).

PLASTER CASTS

Plastic Moulding Compounds

Casts can be produced from ordinary segmented plaster, gelatine, or dibutylphthalate-polyvinylchloride moulds (Stahl, 1956), or by using latex as the impression material (Quinn, 1940), and backing with a plaster jacket (Rau, 1955). As these moulds are in two or more sections, careful keying is necessary to ensure that all parts mate-up tightly together, so that the plaster will not seep through the join and no distortion will appear in the cast. By the use of plastic "hot melt" compounds, one-piece moulds can be produced from almost any fossil however complicated the shape may be. The substance used in the New Zealand Geological Survey is Vinamold, produced by Vinatex Ltd., Devonshire Road, Carshalton, Surrey, England, and sold in New Zealand by R. & E. Tingey & Co. Ltd., Wellington.

Vinamold can be described simply as a "plastic rubber" compound, based on vinyl resins, which when heated is reduced to a liquid that can be poured easly and on setting produces a tough flexible rubbery substance. Vinamold was developed specially for use as a moulding agent for producing high-quality cases.

Vinamold is suitable for casting most fossils, but should be avoided for porous specimens, which it may penetrate too deeply for extraction, and for delicate shells of lamellar structure, which may be damaged by penetration of the moulding compound beneath surface layers. It does not affect glue. Vinamold is available in various grades of hardness, colours denoting the several grades. As all grades can be mixed in varying proportions, a wide range of flexibilities is available. The hardness of the material to be used depends on the nature of the specimen to be cast. As all the grades have a relatively high melting-point, hot water will not affect them. Detergents and soapy water can also be used for cleaning without any injurious effects. The material can be used repeatedly without deterioration. Storage has no effect on the moulds, and dozens of casts of uniform quality can be taken from one mould.

Preparation of the Casting-box

In preparing a fossil specimen for casting, all loose matrix should be removed, the specimen lightly dusted with french chalk, and a suitable casting-box prepared. The box can be a simple construction built up from hard cardboard, stapled at the corners and open at the top. The dimensions of the box depend on the size of the specimen, the grade of Vinamold used, and the rigidity of the mould required.

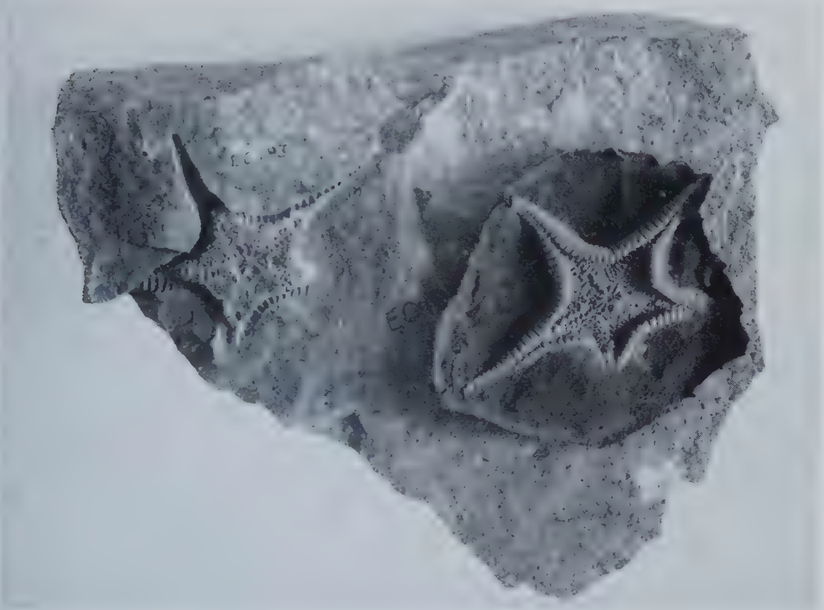

FIG. 1.—Two-tone rubber latex positive of Urutawan (Cretaceous) starfish with original mould.

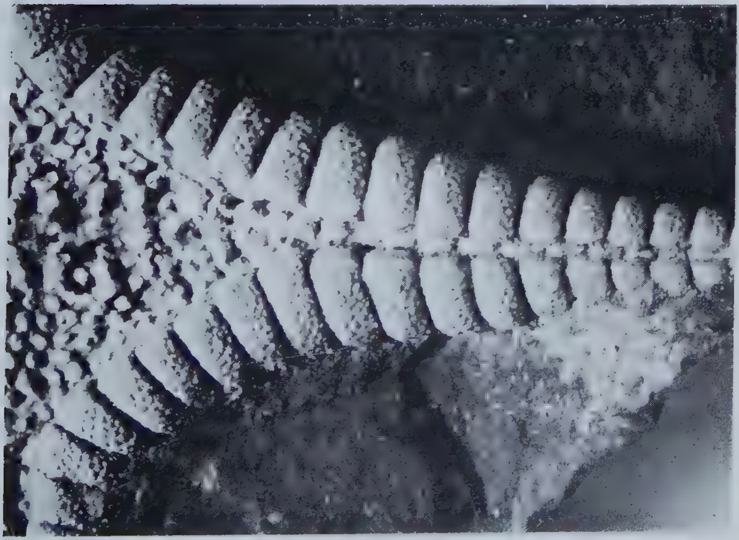

FIG. 2.—Details (× 4) of starfish arm, from rubber latex positive (see Fig. 1).

For fossils an inch or two in diameter, the box should allow about half an inch of air space all round. A fossil exposed on a block of matrix, or any specimen with an area of unimportant matrix on its surface, can be placed in a box for casting using the matrix as the base. As the bottom of the casting-box will later contain the aperture through which the mould is filled with plaster, the fossil should, if possible, be arranged with an unimportant surface downward (Fig. 3). Irregular specimens can be securely positioned by using sealing-wax to attach them firmly to the bottom of the casting-box. Gastropods or ammonites can be arranged to stand on the matrix filling the aperture.

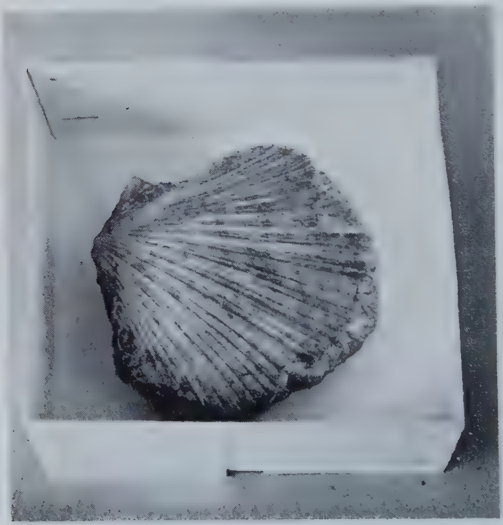

FIG. 3.—Specimen in cardboard box ready for filling with Vinamold.

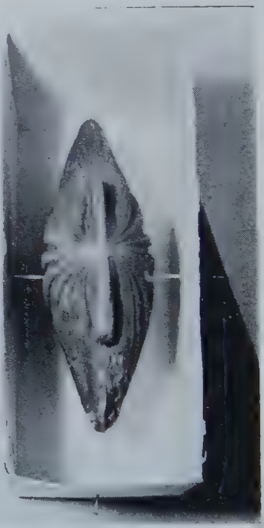

FIG. 4.—Specimen in box supported by pins.

Specimens that are completely free from adhering matrix require careful supporting. If wax is unsuitable to attach the fossil to the bottom of the box, the specimen can be poised by the use of pins. These should be inserted through the sides of the box, with their points touching the fossil so as to hold it in place while the Vinamold is being poured (Fig. 4). The hinge margin of a bivalve fossil should be placed uppermost. The height of the moulding-box should allow for at least a quarter of an inch of Vinamold to cover the specimen, especially as shrinkage during setting leads to a slightly concave surface. A cardboard box is simpler to construct and remove than a wooden case, and serves the purpose equally well. A mould of a small fossil on a large slab of matrix can be taken by building a wall of cardboard around the fossil and fastening the cardboard to the matrix with plaster.

Pouring the Mould

The temperature required for melting Vinamold ranges from 130° C to 170° C, depending on the grade used. A saucepan with a lid, and an electric hotplate, is the simplest equipment for melting the material. As a heat source, a hotplate is preferable to a Bunsen burner, for a single gas jet produces local burning in the mixture owing to heat concentration in a small area. If large moulds are to be made frequently, an oil bath is desirable.

The Vinamold is cut into small pieces to speed up melting. Stirring helps rapid melting, circulates the liquid, tends to equalize any slight differences in temperature throughout the mixture, and reduces the risk of burning. Once the required quantity has been melted and all lumps digested, it can be poured into the casting-box. Further heating will not increase the fluidity, and excessive heating will cause burning. During heating, a lid on the saucepan prevents the escape of unpleasant fumes and loss of volatile constituents. The first indication of burning is the production of extremely acidic fumes and a darkening of the material, with a loss of its elastic properties at a temperature of about 200° C. Burning causes the Vinamold to decompose and harden.

The liquid Vinamold should be poured into the corner of the casting-box in a steady stream, as pauses cause layering in the mould due to surface chilling. Bubbles often rise to the surface of the liquid in the box due to the escape of trapped air from the fossil, but these soon disperse and normally do not affect the mould.

When the mould is completely cool, the cardboard container may be removed. The mould is then inverted and the specimen removed from the base. If necessary, Vinamold can be trimmed away from the opening to allow for ease of removal. For specimens lacking a natural "base" that have been supported by pins during pouring, an opening may be created by slitting the mould. In casting pelecypods, the slit opening is best made to correspond with the junction of the margins of the two valves. By systematically distorting the mould in the fingers, the specimen is loosened and withdrawn. The interior of the mould should be rinsed with water and checked for faithful reproduction of details.

If a fossil occurs as a natural mould, it is often desirable not only to make a plaster replica of the mould, but also to produce a cast of the original specimen. To produce a plaster cast in positive form from a natural mould, two moulds must be made. The ideal combination is the use of liquid rubber latex and Vinamold. The first impression is taken with rubber latex to give a positive. To avoid distortion when the Vinamold is poured on top, the latex should be well strengthened by cotton wool or by a plaster backing. A mould can be taken from this rubber impression in the usual manner. A wall of cardboard can be built around the latex positive, or it can be placed in a box, which is then filled with Vinamold. The latex is unaffected by the hot Vinamold and the resultant mould will be a negative, identical with the original specimen, from which positive plaster casts can be taken.

Casting

The type of plaster used is a rapid-setting, superfine dental plaster. While mixing the plaster, slow careful stirring will prevent the formation of bubbles, thus eliminating holes in the surface of the cast. When the mould is filled, squeezing and movement of the elastic walls will assist any trapped air to rise to the surface and allow the plaster to reach all the extremities of the mould (Fig. 5). After the plaster has set, slight distortion of the mould will release the cast (Fig. 7). A mould can be filled through a slit opening by distorting it. Surplus plaster is squeezed through the opening, and the sides of the slit brought together. When set, such a cast is quite free from attached "pouring-hold plugs" (Fig. 6). These are often only a paper-thin layer which can easily be broken off.

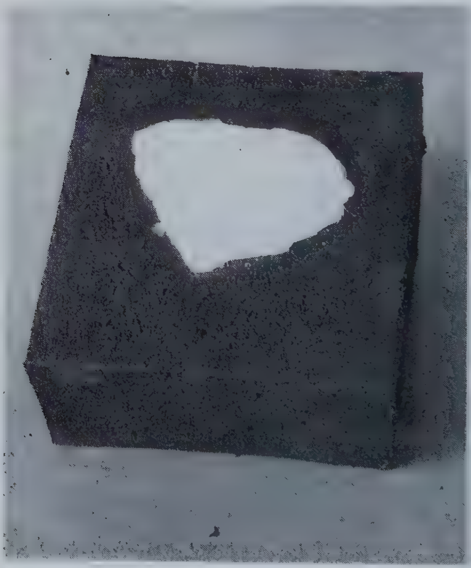

FIG. 5.—Mould filled with plaster.

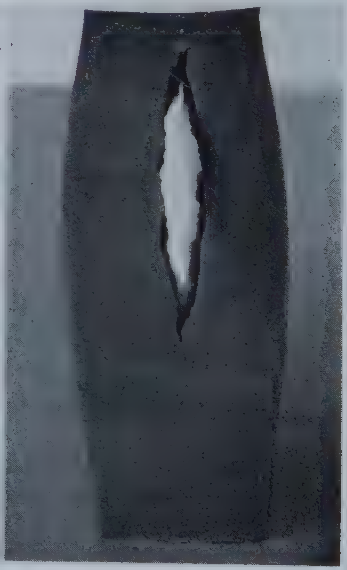

FIG. 6.—Plaster cast being removed from mould through a slit opening. The cut in the mould corresponds with the junction of the two valves.

Support is necessary for large moulds that have only thin walls, to prevent distortion from the weight of the plaster before setting. Where large objects have to be cast and the quantity of Vinamold available is limited, a second identical box may be necessary to support the walls of the mould while the plaster is setting.

Colouring of Casts

Colouring of casts, although not strictly necessary for research work (Fig. 8), gives individuality to the specimens and emphasizes details of sculpture (Fig. 9). When casts have completely dried out, they may be painted with either oil or water colours. Casts painted with water paints require a protective layer of shellac or cellulose paint (Fig. 10).

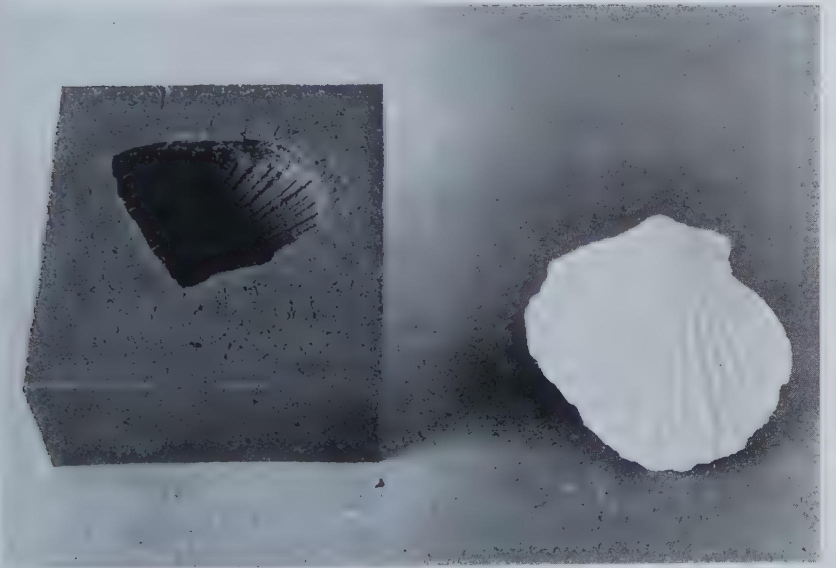

FIG. 7.—Mould and resultant plaster cast.

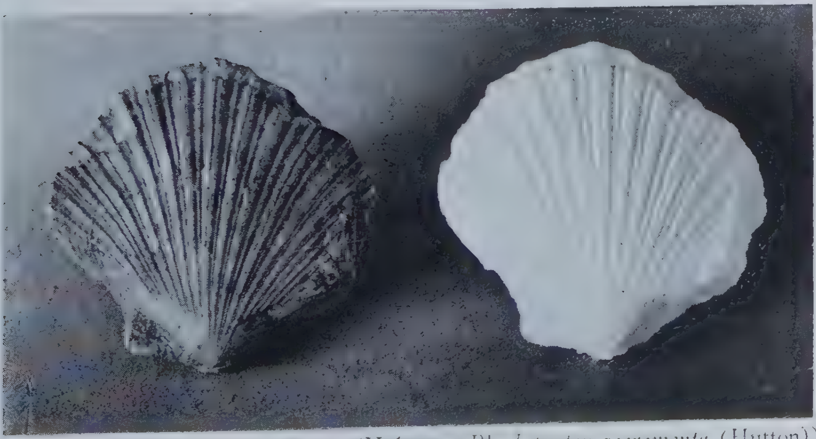

FIG. 8.—Cast and original specimen (Holotype, *Phalopecten accrementa* (Hutton)).

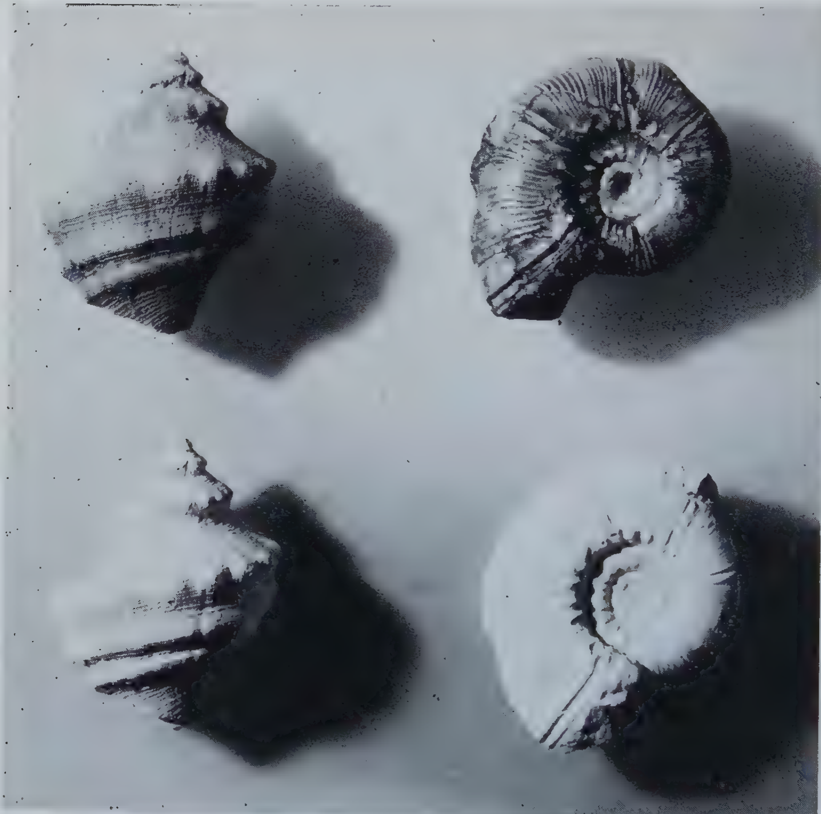

FIG. 9.—Painted and unpainted casts of a gastropod (Paratype of *Taioma tricarinata* Finlay and Marwick) and an ammonite (Holotype of *Madrasites multicostatus* Marshall).

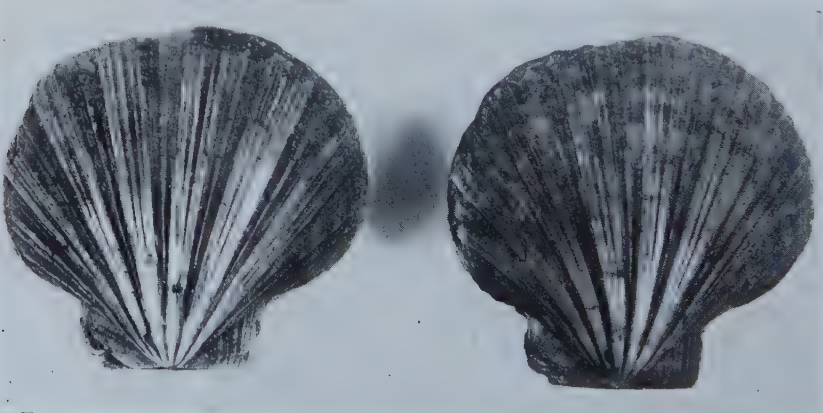

FIG. 10.—Original specimen (Holotype, *Sectipecten wollastoni* Finlay) and painted cast.

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The writer wishes to acknowledge the help given by Mr C. Shaw, and the preliminary work done by Mr G. L. Adkin in the early stages of casting experiments. Vinatex Ltd. and Rubber Distributors Ltd. kindly gave permission to publish details of their products. Special thanks are due to Dr C. A. Fleming for his valuable suggestions and assistance, and to Mr S. N. Beatus for the photographs.

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THE PHASE VELOCITY OF WAVE-COMPONENTS UNDER THE ACTION OF WIND

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Summary

Some observations on the apparent phase velocity of the frequency components in a wave system are described in Part 1. These show that the velocity between two points arranged in the wind direction is higher than that for classical small amplitude waves. Part 2 is a theoretical investigation of the effects of various features occurring in real wave-systems upon the apparent phase velocity of wave components..

Introduction

During the process of the generation of waves by wind there is a complicated system of water motions, in which waves, (mean) currents, and turbulence are all present. In order to investigate the physical processes involved in the transfer of momentum and energy from wind to water, and in particular to the wave-system, it is necessary to know how the components of a spectrum of waves are propagated.

Barber and Ursell (1948) showed that waves which leave a generating area in the ocean and then travel through areas where the "wind is light", behave as a continuous spectrum of component wave trains which travel independently with the classical group velocities appropriate to their periods. These authors measured the time taken for the energy contained by the components of a given frequency to travel from the distant generating area to the recording station. For most of this distance these waves resemble a two-dimensional system of wave motion, with the long crests typical of swell, and with a small steepness.

In this paper observations on the components in a short-crested wave system during an early stage of its generation, are described. The waves were observed on a reservoir with a fetch of 1,200 metres when the wind velocity was about 650 cm/sec. Two wave recorders were placed 346 cm apart in a line along the wind direction. The amplitudes of the Fourier harmonics were found for simultaneous records at each place, and then for their sums with various time lags. In the latter cases the corresponding components are added if the components are in phase, and are subtracted if they are 180° out of phase. Using this technique we may estimate the lag covariance between the components of the two wave records which lie in a restricted range of frequency. An estimate of the phase velocity of each Fourier component from the time required for waves of a given frequency to travel from one recorder to the other is obtained.

N.Z. J. Geol. Geophys. 2 : 66-87.

In Part 2, various factors affecting the observed phase velocities of growing waves are considered. There has been little evidence in the past to show that such wave components act more or less independently and like ideal "classical" waves, whereas there have been suggestions that there are considerable discrepancies in observed wave velocities from those expected from the observed wavelengths. These, however, were based on observations of individual wave crests, and not on any consideration of the spectrum. The shape of an individual wave will bear some relation to the wave-spectrum, and so will its velocity, but until much more is known concerning these relations, no conclusions regarding the phase velocities of the waves or the rate of propagation of wave energy can be drawn. It is hoped that the present work will help in further considerations of these problems.

PART 1—AN OBSERVATION OF THE PHASE VELOCITIES IN A REAL WAVE SYSTEM

THE THEORY CONCERNING THE PHASE VELOCITIES OF THE COMPONENTS IN AN IDEAL WAVE SYSTEM

An Ideal System in Which All Waves Travel in the Same Direction

We suppose first that the wave system is linear and consists of a spectrum of waves such that the energy in any frequency interval is comprised of the sum of many independent, small amplitude, sinusoidal waves all travelling through otherwise still water in one direction θ measured clockwise from a horizontal axis Ox. It is assumed that each frequency component at any point consists of the sum of many "ideal" waves of that frequency, having originated at many points along the line. Let the axis Oy be at right-angles to Ox. Then if we observe these waves at point A with co-ordinates x, y the recorded trace may be represented over a time T sec by a Fourier Series whose n^{th} term is given by

$$h_n = a_n \cos 2\pi\{k_n(x \cos\theta + y \sin\theta) - f_n t + \varphi_n\} \quad \text{............} \quad (1)$$

where a_n is the amplitude and $f_n = n/T$ sec^{-1} is the frequency of the n^{th} harmonic of T, k_n is the reciprocal of the wavelength of an ideal wave of frequency f_n, and the phase φ_n is a random variable. This representation is similar to that used by Rice (1944) to represent currents in electronic valves, and has been used by several authors to represent wave systems. The recorded elevations h'_n at a point B, near A with co-ordinates x', y' at time t' are given by

$$h'_n = b_n \cos 2\pi\{k_n(x' \cos\theta + y' \sin\theta) - f_n t' + \varphi_n\} \quad \text{............} \quad (2)$$

where the phase φ_n is the same as at A and where b_n and a_n are proportional if the wave recorders are linear, or if α is a constant,

$$b_n = \alpha a_n \quad \text{............} \quad (3)$$

The energy contained in each record between the frequencies f_n and $f_n + \triangle f$ is defined by

$$\begin{aligned} B_n{}^2 \triangle f &= \tfrac{1}{2} b_n{}^2 \\ A_n{}^2 \triangle f &= \tfrac{1}{2} a_n{}^2 \end{aligned} \qquad \text{.........} \quad (4)$$

where $\triangle f = 1/T$ sec^{-1} is the frequency interval between adjacent harmonics (e.g., see Rice, 1944). This "energy" is proportional to the potential energy $\frac{1}{4} \rho g c_n{}^2$, where ρ is the water density, g is the acceleration due to gravity, and c_n is the amplitude of the actual wave component.

The n^{th} harmonic component of the sum of the two records has the squared amplitude given by

$$c_n{}^2 = a_n{}^2 + b_n{}^2 + 2a_n b_n \cos 2\pi\{k_n(\xi \cos\theta + \eta \sin\theta) - f_n\tau\} \qquad \text{.........} \quad (5)$$

where $\xi = x' - x$, $\eta = y' - y$, $\tau = t' - t$.

If we write

$$\left.\begin{aligned} R_n{}^2 &= a_n{}^2 + b_n{}^2 \\ D_n{}^2 &= (a_n + b_n)^2 \end{aligned}\right\} \qquad \text{.........} \quad (6)$$

then we have

$$\frac{c_n{}^2 - R_n{}^2}{D_n{}^2 - R_n{}^2} = \cos 2\pi\{k_n(\xi \cos\theta + \eta \sin\theta) - f_n\tau\} \qquad \text{.........} \quad (7)$$

If the ideal wave components all travel along the axis Ox then θ is zero, and (7) becomes

$$\frac{c_n{}^2 - R_n{}^2}{D_n{}^2 - R_n{}^2} = \cos 2\pi(k_n\xi - f_n\tau) \qquad \text{.........} \quad (8)$$

By placing wave recorders at A and B along the direction of wave propagation, then if the real wave system is reasonably described by the representation of this section, the left-hand side of (8) may be computed from the observed Fourier amplitudes. These observed values do not correspond exactly (even to within experimental errors) with values of (8) computed from the right-hand side. A comparison for various frequency intervals is made below, but first we will consider a more complex system in order to examine the effect of waves arriving from more than one direction.

A Complex Three Dimensional System of Ideal Waves

We consider the properties of wind generated waves, and assume that a mean wind velocity exists, on which are superimposed fluctuating velocities whose components vary in space and time. Waves arriving at a point result from the action of the mean plus fluctuating wind components acting at a large number of points and times, and arriving from many directions. The expressions (1) to (3) may now be taken to apply to those wave components arriving from between the directions θ and $\theta + d\theta$. Then, using (3), the relation (4) in this case is

$$B_n^2(\theta)\,.\,\triangle\theta\,.\,\triangle f = \tfrac{1}{2}\,b_n^2 = \tfrac{1}{2}\,\alpha^2\,a_n^2 = \alpha^2\,A_n^2(\theta)\,.\,\triangle\theta\,.\,\triangle f \qquad \text{.........} \quad (9)$$

where B_n^2 and A_n^2 are functions of θ and represent the energies contained at frequency f_n and direction θ, per unit frequency and per unit angle.

Instead of the relation (5), we have

$$c_n^2 = 2\,A_n^2[1 + \alpha^2 + 2\alpha\cos 2\pi(k_n\xi\cos\theta - f_n\tau)]\triangle\theta\,.\,\triangle f \qquad \text{.........} \quad (10)$$

where we have put $\eta = o$, i.e., we take AB along the wind direction.

In practice we do not observe a_n^2, b_n^2, c_n^2 but the sum due to components near each frequency arriving from all possible directions. Thus, the observed amplitudes squared for the wave records taken at A and B are

$$a'^2_n = 2\int_{-\pi}^{\pi} A_n^2(\theta)\,d\theta\,.\,\triangle f$$

and

$$b'^2_n = \alpha^2\,a'^2_n \qquad \text{.........} \quad (11)$$

while that for the sum is (from (10))

$$c'^2_n = a'^2_n(1 + \alpha^2) + 4\alpha\int_{-\pi}^{\pi} A_n^2(\theta)\cos 2\pi(k_n\xi\cos\theta - f_n\tau)\,d\theta\,.\,\triangle f. \qquad \text{.........} \quad (12)$$

Thus if

$$R_n^2 = a'^2_n + b'^2_n$$
$$D_n^2 = (a'_n + b'_n)^2$$

instead of (7) we may write:

$$\frac{c'^2_n - R_n^2}{D_n^2 - R_n^2} = \frac{\int_{-\pi}^{\pi} A_n^2 \cos 2\pi(k_n \xi \cos\theta - f_n \tau)\, d\theta}{\int_{-\pi}^{\pi} A_n^2(\theta)\,.\,d\theta.} \qquad \text{............ (13)}$$

It is seen that for such a wave system the function calculated from the observed coefficients is a weighted mean of the cosine function in relation (8) for unidirectional waves where the weighting function is the energy of the waves arriving from the direction θ at the frequency f_n. This is used in Part 2, to assess the effect of the multidirectional nature of the waves.

The Techniques Used For Observation and Fourier Analysis of the Wave System

Capacitance-wire recorders similar to those described by Tucker and Charnock (1955), were used. The measuring head consists of a polythene-insulated copper wire placed vertically through the water surface. The inner conductor and the surrounding water form the plates of a capacitor which is part of a tuned circuit of an oscillator whose output is frequency modulated according to the changes of water level. This signal is passed through a frequency discriminator similar in principle to that used in F.M. radio receivers, and the output, which is linear with changes of water level, is fed to a mirror galvonometer. The waves at two recorders 346 cm apart fixed on a rigid horizontal arm to a pole in the water were recorded photographically on the same chart together with time marks every two seconds.

The arm supporting the recorders was placed as nearly as possible in the mean direction of the wind by noting the direction of a wind vane, the general direction of wave crests and the drift current. An error of 14° in this direction will cause a difference of less than 3% in the distances between the wave recorders and this distance projected along the true mean wind direction.

The wave records were produced as a black line on white paper. These were curve-followed and converted to half-black and half-white records using an improved version of the photo-electric curve-follower described by Tucker and Collins (1947).

The black and white records representing the waves occurring simultaneously at each of the two sites, were both analysed on the Fourier analyser described by Barber, Ursell, Darbyshire, and Tucker (1946). The record is fitted round the circumference of a wheel which is rotated at high speed and then allowed to slow down. Light reflected from the white part of the record falls on to a photo-electric cell in whose output the wave record is reproduced electrically. This signal is fed to a high-Q galvanometer which resonates whenever the wheel speed is such that a harmonic of the length of the record reaches the resonant frequency. Thus, the output is a series of peaks proportional to the Fourier amplitudes.

For the particular occasion chosen for the present experiment, the overall frequency response was such that the analysis is essentially linear for all frequencies less than three cycles per second (corresponding to ideal waves of lengths greater than 16 cm). Simultaneous records from each site (of duration 85·5 sec.) and then their sum (i.e. both records placed on the wheel), were analysed. The downwind wave record was placed at successively later times relative to the upwind record, in steps of 0·25 sec. The actual values of the 18 lags were found by drawing a series of lines across the black and white records for both the greatest and least time lags and then identifying the particular waves cut by these lines on the original records. The error involved in this lag measurement is about 0·1 sec.

Small variations of the recording speed during the observations of the waves have the same effect on both wave records, and merely cause a slight spreading of the energy at any given frequency through nearby frequencies. The independent variations of speed during the process of curve-following each record may produce a further blurring which is reduced to a minimum by choosing a pair of black-and-white records for which there was no measurable difference in total length (equivalent to 85·5 sec.). The total effects of these variations are much less than those due to the filter effect of the Fourier analysis process itself.

THE DATA OF THE EXPERIMENT

Observed values of the expression (8) were computed from the observed Fourier amplitudes. Scatter in these values may be introduced during the preparation of the records for analysis. The Fourier filter responds to frequencies near that of each harmonic (see Pierson, (1954) or Barber and Ursell (1948)), and this also results in the spreading of the energy present in a record at a given frequency through several harmonics. In order to reduce the scatter due to these effects and to the natural variance of the data, the energy contained in a frequency band equivalent to that covered by five adjacent harmonics was used. Thus, the values of a_n^2, b_n^2, and c_n^2 corresponding to (4) and (5) were the sums of the observed squared amplitudes of five consecutive harmonics. The frequency interval covered is $5\triangle f = 5 (1/85{\cdot}5) = 0{\cdot}058\ \mathrm{sec}^{-1}$, where the duration T of each record was

85·5 sec. Then the values computed from the left-hand side of (8) apply to this larger frequency interval, and are an average for all components contained within this interval.

If the values of a given group are plotted against the lag τ, the curve obtained by joining the points in a reasonably smooth curve closely resembles a sinusoidal curve. The observed curves are shown in Fig. 1, where the zero for each curve is the horizontal line at the mid-frequency of the group, f_n, and the lines at the adjacent frequencies have the values ± 1. These curves may be compared with those obtained from the right-hand side of relation (8) where $\xi = 346$ cm and $k_n = 2\pi f_n/g$. A displacement $-\delta\tau$ of the observed curve to the left implies that the wave-components travel between the wave recorders with an apparent phase velocity greater than that of ideal small-amplitude waves of the same frequency.

If the observed data are fitted to a curve $(\alpha_n + \beta_n)^{\frac{1}{2}} \cos\{2\pi(k_n\xi - f_n\tau) + \delta_n\}$ then δ_n is the lag by which the observed data differ from that cosine curve expected for an ideal wave component. If $x(f_n, \tau_i)$ is an observed value of (8) then we may estimate the time lag difference $\delta\tau$ from the relation

$$-\delta_n = 2\pi f_n \delta\tau = \tan^{-1}\frac{\beta_n}{\alpha_n} \qquad \text{..........} \quad (14)$$

where

$$\begin{matrix}\alpha_n \\ \beta_n\end{matrix} = 2/18 \sum_{\tau_i} x(f_n, \tau_i) \begin{matrix}\cos \\ \sin\end{matrix}\left(\frac{4\pi^2 f_n^2}{g}\xi - 2\pi f_n \tau_i\right) .$$

The computed values of $\delta\tau$ are given in Table 1. In Fig. 1 the straight line is the time T_n for ideal waves of frequency f_n to travel the distance ξ and the shaded area shows the computed values of $\delta\tau$.

At the lowest frequencies in the spectrum the waves may be expected to travel most nearly at the classical wave velocities. Thus, these waves and those up to the frequency 1·08 sec^{-1} may be associated in phase with the ideal waves. However, for higher frequencies it was not possible to recognize a pattern in the lag differences, a matter which is referred to in Part 2.

Now if T_n is the time taken by ideal waves of frequency f_n to travel through the distance ξ, ξ/T_n is the phase velocity of these ideal waves, and if $\tau_n = T_n + \delta\tau$, then ξ/τ_n is the observed phase velocity of the real waves. Thus the percentage difference of the real wave velocity to the ideal wave velocity is:

$$100\frac{\xi/\tau_n - \xi/T_n}{\xi/T_n} = 100\frac{T_n - \tau_n}{\tau_n} = 100\frac{-\delta\tau}{\tau_n}\% .$$

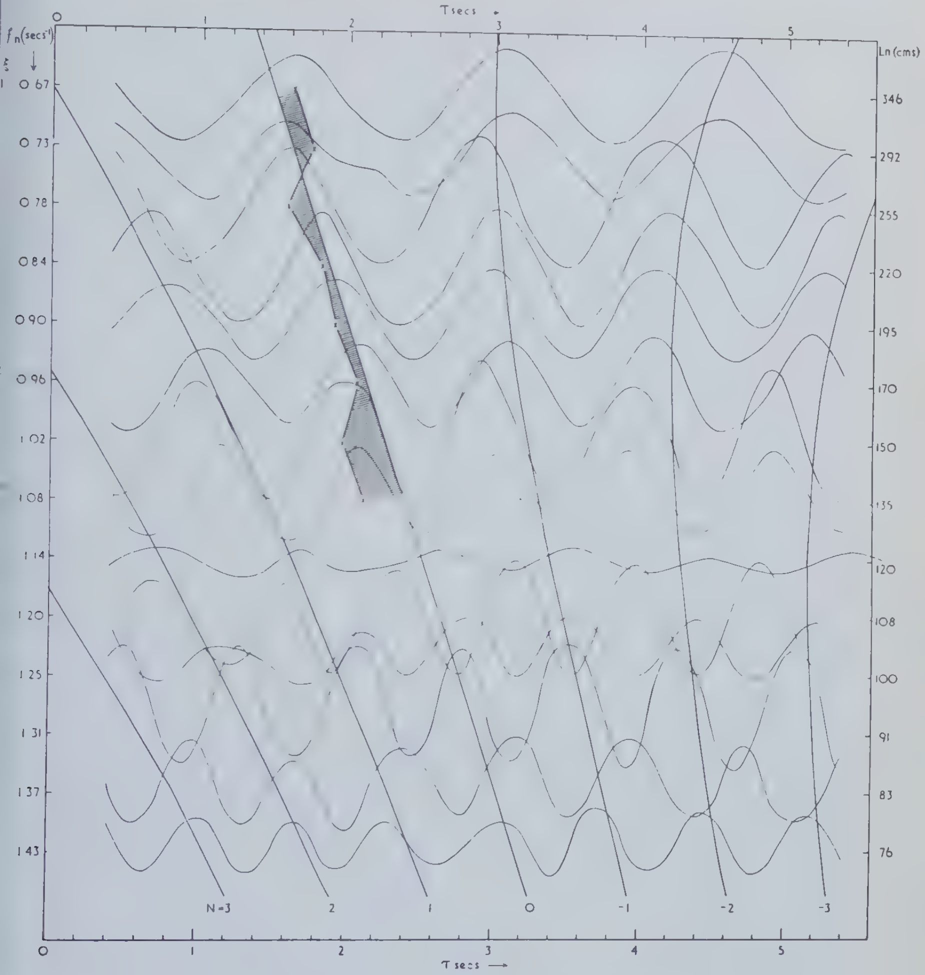


FIG. 1.—Curves obtained from the computed values of the correlation coefficients (relation (8)) for varying lags τ. The downward curves show the positions of the maxima of $\cos 2\pi(k_n\xi - f_n\tau)$ and are computed from $k_n\xi - f_n\tau = N$. The straight line at $N = 0$ represents the time taken by ideal waves to travel between the recorders and the shaded portion indicates the difference in travel time for the observed waves.

This is also shown in Table 1, together with the percentage of the total energy ($n = 52$ to $n = 127$) in each frequency interval, and the cumulative total of this ratio.

TABLE 1.

f_n sec^{-1}	0·67	0·73	0·78	0·84	0·90	0·96	1·02	1·08
$n =$	57	62	67	72	77	82	87	92
δr sec	0·13	0·14	−0·16	−0·06	−0·10	−0·08	−0·30	−0·29
T_n sec	1·48	1·61	1·73	1·86	1·99	2·12	2·25	2·38
τ_n sec	1·61	1·75	1·57	1·80	1·89	2·04	1·95	2·09
$-\delta\tau/\tau_n \times 100\%$	−8·1	−8·0	10·2	3·3	5·3	3·9	15·4	13·9
% of energy	4	15	15	30	6	11	4	4
Cumulative energy %	6	21	36	66	72	83	87	91
$\sqrt{\alpha_n^2 + \beta_n^2}$	0·59	0·43	1·23	0·96	0·66	0·68	1·21	0·66
$\delta\tau'$ sec	0·08	0·12	−0·15	−0·05	−0·10	−0·08	−0·28	−0·30
$\sigma_{\delta\tau'}$ sec	0·04	0·10	0·02	0·07	0·04	0·03	0·03	0·04

AMENDED TABLE

$\delta\tau''$ sec	0·00	0·01	−0·29	−0·19	−0·23	−0·21	−0·43	−0·42
τ_n'' sec	1·48	1·62	1·44	1·67	1·76	1·91	1·82	1·96
$-\delta\tau''/\tau_n'' \times$ 100%	−0·0	−0·1	20·1	11·4	13·0	11·0	23·6	21·4

The values of the observed coefficients $\sqrt{\alpha_n^2 + \beta_n^2}$ (see relation 14) are also given in Table 1. For a single ideal wave-component this amplitude is theoretically equal to one. For a multi-directional wave-system the amplitudes expected when recorders are placed at various spacings ξ, and for several forms of the directional spectrum $A^2(\theta)$ are shown in Fig. 4. The figures in Table 1 indicate that there is considerable scatter in the estimated values. This may arise quite largely from the process of making three separate operations to find a_n, b_n, and c_n in relations (4) to (8). That this is so is indicated by the fact that some curves in Fig. 1 are displaced from their zero, e.g., $n =$ 62, 67, and 77. (This computation in (14) was made with respect to an estimated zero for each curve). The mean value of the observed amplitudes is 0·80 with a root mean square deviation of 0·28. An amplitude even as low as 0·52 may be seen from Fig. 4, for values of $2\pi k\xi$ between 6 and 16, to indicate that $A^2(\theta)$ decreases no less rapidly, as θ increases, than $\cos\theta$ or perhaps $\cos^2\theta$. However, most curves in Fig. 1 closely resemble sine curves and the estimate of phase will not be seriously affected by errors causing scatter in the amplitudes.

An estimate of the standard deviation of the computed phase lags may not be readily made directly from the data. However, if we estimate the positions of the maxima of the curves in Fig. 1, the deviations from the downwards sloping lines are also estimates of $\delta\tau$ and the means $\delta\tau'$ for each n are also given in Table 1 together with their root mean square deviations $\sigma_{\delta\tau'}$. The root mean square value of this last row is 0·053 and the absolute error in the estimate of the lags τ may be as high as 0·1 sec. Thus, it is seen that the values of $\delta\tau$ and $\delta\tau'$ at $n = 87$ and 92 indicate that the apparent phase velocity of the real wave components is higher than that of ideal waves. There is also possibly an increase in the case of $n = 67$ where the value of $\sigma_{\delta_T'}$ is only $0 \cdot 02$ sec, however a single value of this standard deviation may be well in error due to its method of estimation, whereas the estimate $0 \cdot 053$ sec will be quite reliable.

In Part 2 it is argued that the longest waves will travel nearly at the velocity for ideal waves. We note here that there is no mechanism which acts to slow down the waves (except possibly turbulence, which should have its least effect on the longest waves). Thus, we expect that the values of δ_T for these longest waves should be zero within $\pm 0 \cdot 05$ sec. Also the error of about $0 \cdot 1$ sec involved in the original estimate of the lags must act in the same direction on all estimates of δ_T. The amended table shows revised values δ_T'' if $-0 \cdot 13$ sec (reducing δ_T for $n = 57$ and 62 to nearly zero) is added to all values of δ_T, and the revised values of the percentage apparent phase velocity increases are greater. It may be expected that the real velocity increases will be somewhere between those indicated in the original and amended tables.

We return to these estimates in the discussion at the end of Part 2.

PART 2—A DISCUSSION ON SOME FEATURES THAT MAY AFFECT THE PHASE VELOCITIES OF REAL WAVES

It is evident from the work described in Part 1 that some wave components travel with phase velocities greater than those expected for ideal waves. We will attempt to estimate the possible effect on their velocity of some features existing in the presence of real waves. These are:

(a) the fact that real wave-components approach a point from more than one direction,

(b) the water-drift velocities on which the wave system is superimposed,

(c) the wave steepness and the shape of the waves,

(d) the group-nature of the wave-components,

(e) the fact some wave-components may lose their identity in travelling between the recorders, i.e., these waves may be uncorrelated between the sites *A* and *B*,

(f) wind action on the wave system itself,

(g) turbulence in the water.

Obviously not all of these are independent in the complex non-linear wave system. However, here an attempt is made to assess the effect of each separately.

(a) The effect of the multi-directional nature of waves is illustrated by the relation (13) in Part 1. It is assumed that the energy per unit of angle is greatest in the wind direction ($\theta = 0$) and that $A^2(\theta)$ is symmetrical about this direction and always decreases as θ increases from zero. This is in accordance with similar properties of the generating wind. We can calculate values of the numerator on the right of the relation (13) expressed as,

$$C \cos 2\pi f\tau + S \sin 2\pi f\tau \qquad \text{..........} \quad (15)$$

by assuming various functional shapes for the directional spectrum $A^2(\theta)$. In this expression

$$\begin{matrix} S \\ C \end{matrix} = 2\int_0^{\phi} A^2(\theta) \begin{matrix} \sin \\ \cos \end{matrix} (2\pi k\xi \cos \theta)\, d\theta \; . \qquad \text{..........} \quad (16)$$

are functions of ϕ, $A^2(\theta)$ falling to zero, perhaps rapidly, at $\theta = \phi$. The amplitude of the function in (13) for various forms of $A^2(\theta)$ is shown in Fig. (4) as a function of $2\pi k\xi$. In order to avoid serious difficulties in this analysis we further suppose that ϕ is always less than $\pi/2$ radians. Extreme values of the expression (15) as τ varies are found when

$$\tan 2\pi f\tau = \frac{S}{C}, \text{ or } 2\pi f\tau = \tan^{-1}\frac{S}{C} + m\pi \qquad \text{..........} \quad (17)$$

where m is an integer. For some value of m, τ is the time taken for waves of frequency f to travel the distance ξ. Differentiating $f\tau$ with respect to ϕ we find extreme values of $f\tau$ in (17) when

$$\tan (2\pi k\xi \cos \phi) = \frac{S}{C} = \tan 2\pi f\tau \; .$$

The solution:

$$k\xi \cos \phi = f\tau - \tfrac{1}{2} \qquad \text{..........} \quad (18)$$

for $\phi < \pi/2$ may be shown to give that minimum of $f\tau$ as a function of ϕ, which corresponds to a maximum in the apparent phase velocity of the waves of frequency f, when the present method of analysis is used and the distance between the wave recorders is ξ. The solution (18) is indeterminate, but is useful in following considerations.

We note that the ratio $k\xi/f\tau$ is the ratio of the apparent phase velocity (ξ/τ) to the classical phase velocity (f/k) for waves of frequency f.

If $\phi = \pi/2$, values of multiples of S and C may be read from suitable tables, using the following scheme.

$$\begin{aligned} A^2(\theta) &= 1 & 2/\pi . S &= \Omega_0(2\pi k\xi) \\ &= \cos\theta & &= J_1(2\pi k\xi) \\ &= \cos^2\theta & &= \tfrac{1}{2}\left[S_0(2\pi k\xi) - \frac{S_1(2\pi k\xi)}{2\pi k\xi} \right] \end{aligned}$$

$$\begin{aligned} A^2(\theta) &= 1 & 2/\pi . C &= J_0(2\pi k\xi) \\ &= \cos\theta & &= -\,\Omega_1(2\pi k\xi) \\ &= \cos^2\theta. & &= \tfrac{1}{2}\left[J_0(2\pi k\xi) - \frac{J_1(2\pi k\xi)}{2\pi k\xi} \right] \end{aligned}$$

Here, $J_n(x)$ is the Bessel function of order n, $\Omega_n(x)$ is the Lommel-Weber function of order n, and $S_n(x)$ is the Struve function of order n (e.g. see "Tables of Functions", Jahnke and Emde, pp. 150 and 211). Values for $2\pi f_T$ may be found from the relation (17) and the ratios $k\xi/f_T$ are plotted in Fig. 2 against the abscissa $2\pi k\xi$. This shows how the ratio of the apparent to the classical phase velocity depends on the distance ξ between the recording stations for various forms of the directional spectrum .The crosses joined by the dashed line are particular solutions (obtained in the next paragraph) of the equation (18) when $A^2(\theta) = 1$, and represent maximum apparent velocities.

The dependence of the velocity ratios on the angular width ϕ is illustrated in Fig. 3. The solid curves represent the cases for which $A^2(\theta) = 1$ for θ between 0 and ϕ, and otherwise zero. S and C were calculated by integration of (16) at 1° intervals of θ for the cases $k\xi = 1, 1{\cdot}67, 2$, and 3. As a check, figures obtained for $\phi = \pi/2$ differ from the corresponding functions $J_0(2\pi k\xi)$ and $\Omega_0(2\pi k\xi)$ by less than $0{\cdot}001$. Also the maxima of $k\xi/f_T$ found using relation (17) by the integrations, occur at a value of ϕ agreeing to within 30 minutes of angle with that calculated by substituting the value of f_T in (18). These maximum values are plotted in Fig. 2 as crosses, together with a point for which $\phi = \pi/2$, where $f_T = 0{\cdot}5$ from (18) and $2\pi k\xi = 4{\cdot}33$ ($k\xi = 0{\cdot}69$) at the value where $\Omega_0(2\pi k\xi) = 0$ (see the scheme above). A further point appropriate to $2\pi k\xi = 0$ follows readily from the considerations of the next paragraph. It is noted that these points, joined by the dashed curve in Fig. 2, are upper limits of the velocity ratio, given $k\xi$, for all values of ϕ when $A^2(\theta) = 1$, and may possibly be extreme values for all forms of $A^2(\theta)$ under the conditions of this section.

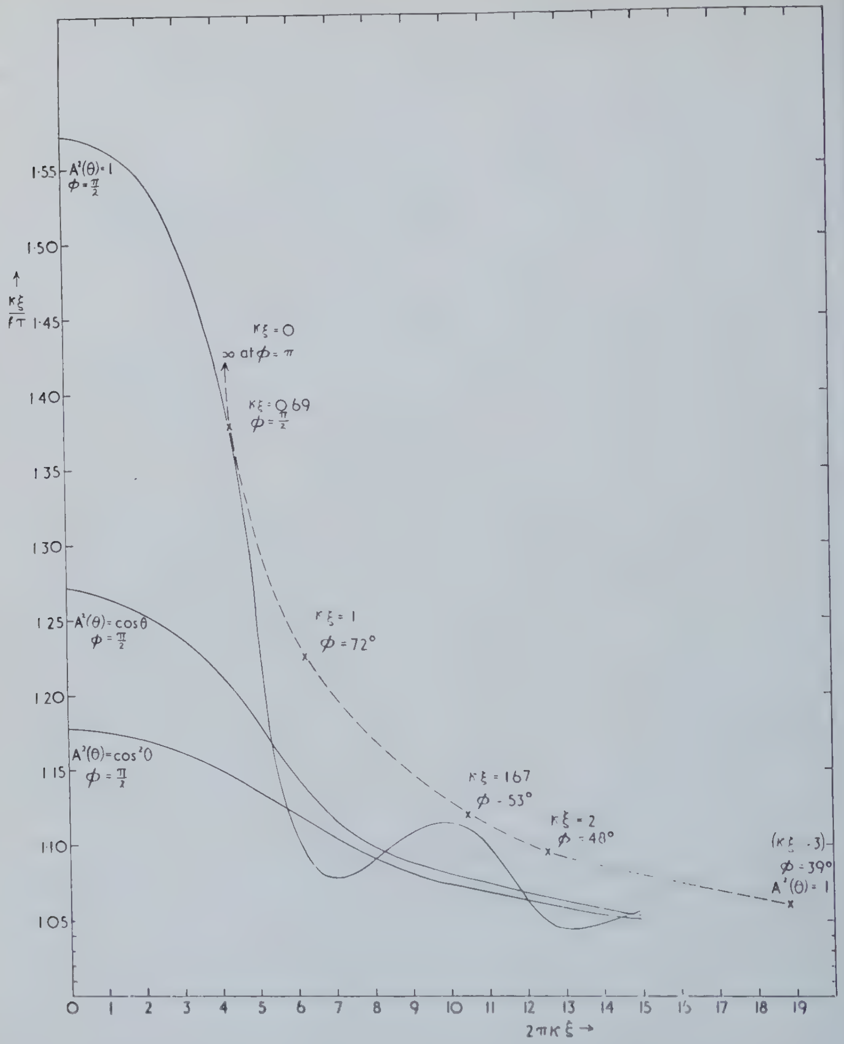
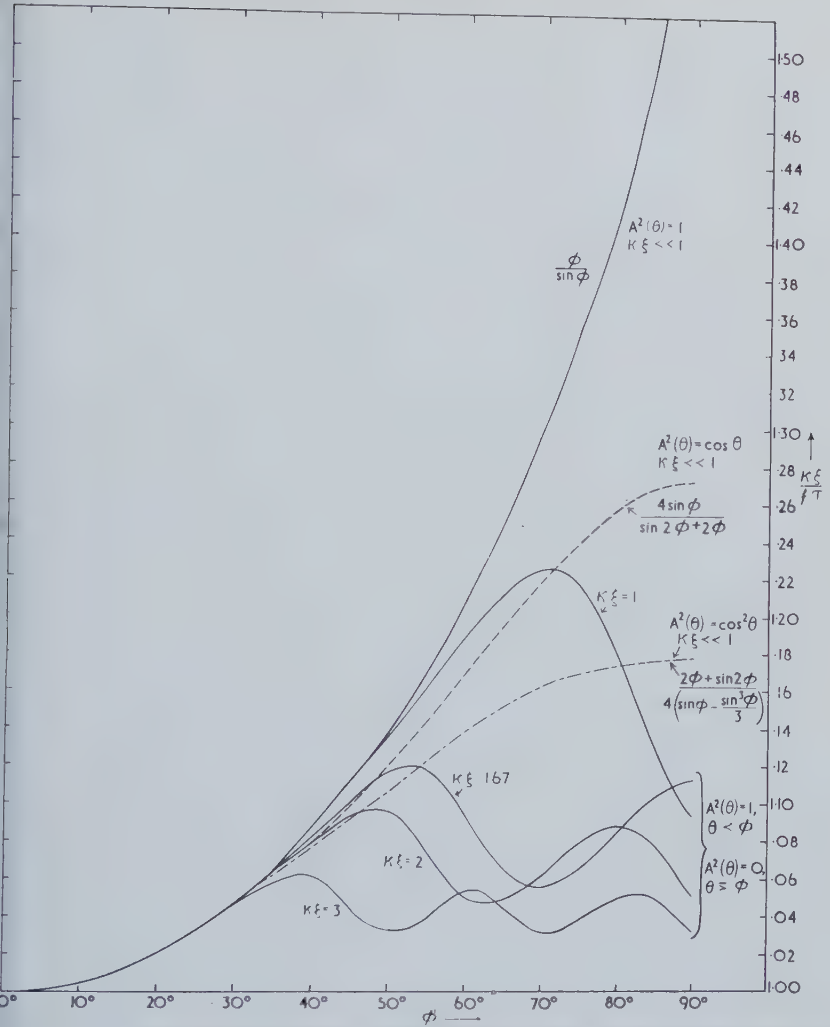


FIG. 2.—The ratio of the apparent phase velocity of multi-directional waves to the classical phase velocity, as a function of the separation of the recorders.

For very small values of $k\xi$, the functions S and C may be integrated to give approximately $2\pi f\tau = S/C$ in place of (17). The ratios $k\xi/f\tau$ found in this manner are shown by curves in Fig. 3 along with the appropriate approximating functions. It is noted that for $A^2(\theta) = 1$, the ratio becomes infinite at $\theta = \pi$ when $k\xi$ is very small. This is the case in which waves approach with equal intensity (for a long time average) from all directions, but we are not concerned with such waves in this paper.



IG. 3.—The ratio of the apparent phase velcity to the classical phase velocity as a function of the width of the spectrum.

From Fig. 3 we see that the increase of the apparent phase velocity is small ($< 5\%$) if the width of the spectrum is less than 30°. It is reasonable to expect that the energy of the longest wave-components in the spectrum progresses mainly in the wind direction since they can be produced only by the breaking of shorter waves, or by the strongest total wind components, which both act mainly along this direction. Shorter wavelength components will be much more affected by the transverse fluctuating wind components and may be expected to have a wider directional spectrum.

The general shape of these curves is very interesting, and may be explained, crudely, as follows: The analysis provides the correlation of the effect of the wave components near a given frequency at each of two points a distance ξ apart lying along the wind direction. This correlation has a maximum at $\xi = 0$, $\tau = 0$ which effectively travels along the direction of increasing ξ with a diminishing amplitude (see Fig. 4). The contribution by components travelling near the angle θ progresses with the velocity $f/k \cos\theta$. The contributions of all components add to the maximum and each plays a share in determining the higher velocity for sufficiently small values for ξ. (See the curves for small values of ξ in both Figs 2 and 3.) At a given value of ξ the velocity at first increases with increasing ϕ, but the contribution from a component at a sufficiently high value of θ runs ahead of the main maximum which is composed of contributions from lesser values of θ, and at still greater angles the principal maximum is reinforced by the following "wave" of this component (e.g. see curves for constant $k\xi$ in Fig. 3). For a given width of spectrum, a similar process occurs in that as ξ increases the contributions from components at a wide angle outstrip the main maximum whose apparent velocity decreases, but at a

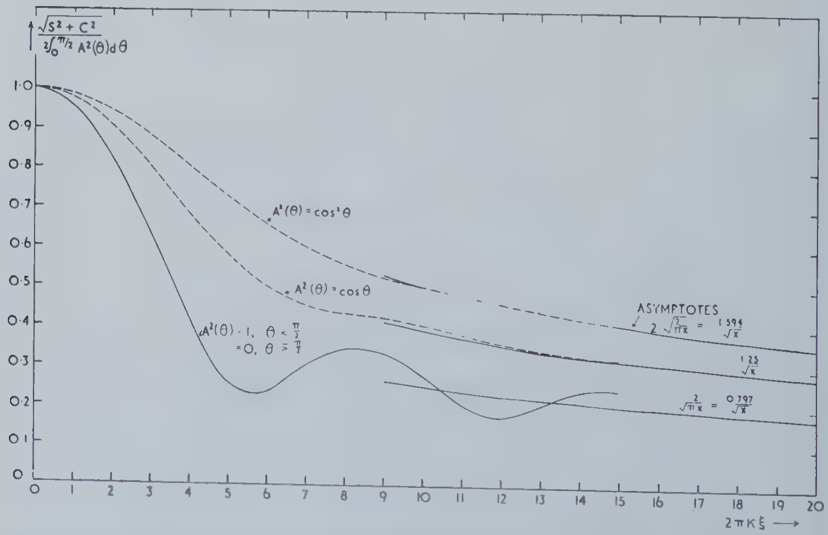


FIG. 4.—The amplitude of the correlation function (see relations (13) and (16):

$$\frac{S\cdot\sin 2\pi f\tau + C\cdot\cos 2\pi f\tau}{2\int_0^{\pi/2} A^2(\theta)\cdot d\theta}$$

still greater ξ (e.g. see curve for $A^2(\theta) = 1$, $\phi = \pi/2$ Fig. 2) the next "crest" of this "faster" component is catching up and the apparent velocity increases. For large values of ξ only those components arriving from small angles of θ contribute to that maximum originally recognized at $\xi = 0$, $\tau = 0$, however, at these separations of several wavelengths, it may be difficult to distinguish this maximum.

(b) The wave-motion system exists near the surface of a mass of water whose near surface layers are also moving in the wind direction. This latter motion has a mean velocity which is greatest at the surface and decreases with depth. Now in an ideal wave system the velocity and pressure components have mean amplitudes decreasing exponentially from the surface (the amplitude of the velocity at depths of one quarter and one half a wavelength for each component are about 21% and 4·3%, respectively, of the surface value) and the physical action of each component depends on the composite of its properties at all depths. It may be expected, then, that very short, high frequency components will have an additional mean velocity in the wind direction almost equal to the surface drift velocity, and that the additional velocity will decrease with increasing wavelengths. Abdullah (1949) and Thompson (1949) have shown that the velocity increase for small waves is equal or less than the surface drift velocity.

Under conditions very similar to those prevailing when the observations described in this paper were made, the surface drift velocity measured by the rate of drift of almost neutrally buoyant spheres from one to three inches in diameter was about 2·5% of the wind velocity at 10 metres. This compares with the figure of about 3·3% usually accepted for the surface drift at longer fetches. We can then set an upper limit to the increment of velocity for wave components, of various frequencies. Table 3 shows the percentage ratio of 16 cm/sec. which is 2·5% of the wind velocity of 650 cm/sec, to the phase velocities of ideal waves for various frequencies.

TABLE 3

n	57	67	77	87	97	107	117
f_n sec—1	0·67	0·78	0·90	1·02	1·14	1·25	1·37
V_n cm/sec	230	200	175	155	135	125	115
$16/V_n \times 100\%$	7	8	9	10·5	12	13	14

These figures will give a considerable overestimate of the increments at the lower frequencies $f_n < 0{\cdot}78\ \text{sec}^{-1}$ (wavelengths greater than 255 cm) where the wave action is still considerable at 1 metre, while for the higher frequencies $f_n > 1{\cdot}14$ (wavelengths less than 120 cm) the wave motion is negligible at a depth of 50 cm and the figures may be nearly correct.

It is remarked that at much greater fetches the wave spectrum moves towards much lower frequencies and for well developed ocean wave systems only the highest frequencies containing an insignificant proportion of the total energy will be significantly influenced by water drift.

(c) Michell (1893) showed that regular two-dimensional waves of finite height, have a maximum possible steepness, i.e. ratio of crest-to-trough height to wavelength, of about $1/7$, and the velocity of these waves is about 10% greater than that of small amplitude waves. In section 3 we considered waves of infinitesimally small amplitudes and of many frequencies, which may be compounded linearly to represent a spectrum. Waves of finite height do not act independently and cannot be summed or integrated in a similar manner, but it is interesting to examine the effect of "steepness" of the wave components in the observed spectra of real waves.

A large number of observations of waves were made on the same reservoir and under conditions extending about those prevailing for the observation described in this paper. These observations are described in a paper in preparation for publication. It is shown that at frequencies higher than that at which the maximum energy density occurs in a spectrum, the density at frequency f is given approximately by

$$F(f) \approx \frac{8 \cdot 5}{f^5}$$

(cm^2 per unit frequency). If we let λ cm be the wavelength and $L(\pi)$ the energy per unit of increment of the natural logarithm of λ, then

$$L(\lambda) d(\log \lambda) = L(\lambda) \frac{d\lambda}{\lambda} = F(f) . \left| df \right| .$$

If the wave components act to a first approximation like ideal small amplitude waves, $\lambda = g/2\pi f^2$ and we find

$$L(\lambda) = \tfrac{1}{2} f . F(f) .$$

If $h(\lambda)$ is the height of a sine wave containing the same energy as $L(\lambda)$, then we will define the “steepness’ of wave components of length λ, as $\{h(\lambda)\}/\lambda$. We have,

$$\frac{h^2}{\lambda^2}=\frac{8L(\lambda)}{\lambda^2}=4\left(\frac{2\pi}{g}\right)^2 f^5.F(f)\approx 4\left(\frac{2\pi}{g}\right)^2 8\cdot 5$$

thus

$$\frac{h}{\lambda}\approx\frac{1}{27}=\text{constant.}$$

Now it is not at all obvious how we may best define a “steepness” for the components of a spectrum, to concur with the usual definition. We note that $L(\lambda)$ is the energy per unit of the proportional increment of wavelength, i.e. of $d\lambda/\lambda$ and this is at least a reasonable figure on which to base a representation of the “steepness” of the components. Moreover, the spectrum $L(\lambda)$ is proportional to the spectra defined from the energy per unit increments of the logarithms of frequency f, period $1/f$ or wave-number ($k=1/\lambda$), and the constants of proportion involved are respectively (to multiply $L(\lambda)$), $\frac{1}{2}$, $\frac{1}{2}$, and 1. Thus the “steepness” derived above is greater than one derived perhaps less reasonably from these latter spectra. In any case, a remarkable feature of such a representation is that the “steepness” is nearly constant for frequencies higher than that at which the maximum power density occurs. The steepness at lower frequencies, or longer wavelengths is evidently very much less.

The increments of velocity for waves of finite height of steepness 1/10, 1/15, 1/20, and 1/25 are respectively 5%, 2·8%, 1·3%, and 0·8% of the velocity of ideal waves of the same wavelengths. It would seem then that the higher frequency wave-components reach an equilibrium stage at which the effect of steepness on their velocities is negligible.

(d) and (e) In part (a) of this section we referred to the role of the fluctuating wind components in generating the wave system. It is evident that this role will cause the components near a given frequency to arrive at a point as a series of groups of waves. Moreover, in such a process of wave generation the phase of one group is independent from that of another. This is true in space as well as in time, thus as the distance ξ between the wave-recorders represents an increasing number of wavelengths, i.e. as the frequency increases, the Fourier analysis method of estimating the correlation coefficients (using relation (8)) will have an increasing scatter, and a space-time correlation coefficient tends to zero as ξ or τ increases. In Fig. 1 the scatter in the amplitudes for frequencies greater than $1\cdot 08\ \text{sec}^{-1}$ is obvious. Of more importance here is the increasing scatter in the phase of the curve calculated from (8), and the low energy content of these higher frequency waves. These facts are probably responsible for the lack of any recognizable pattern in the phases of the curves in Fig. 1 with frequencies greater than $1 \cdot 08 \text{ sec}^{-1}$ and there is no obvious lag on which to base a calculated lag discrepancy.

(f) and (g) There is little than can be said on the possible effects of wind action in accelerating a "wave" as a whole and the effects on its components, or the effect of turbulence in the water on the phase velocity of the wave components. The latter is responsible partly for the transfer of energy through the wave-spectrum towards lower frequencies and partly for a loss of energy from the wave system as a dissipative effect. This latter dissipative process indicates that turbulence may act to slow down a wave system and perhaps its components. If so, it is alone in this as all other features apparently increase the phase velocities. We can only assume here that in neither role does turbulence significantly affect the phase velocity of particular wave-lengths, but it is likely that its influence is greatest on the higher frequency components.

The longest, least steep, wave components are likely to be least affected by either wind or turbulence.

DISCUSSION AND CONCLUSIONS

The factors discussed under the above headings are seen to affect the wave-system in a very complicated manner. It seems fairly certain, however, that the longest waves in the spectrum act nearly independently and will travel nearly like classical small waves. Thus, in Table 1 amended figures for the phase velocity increases are given in the last row, based on the assumption that the observed velocity increment for the harmonics $n = 57$, and $n = 62$ should be nearly zero, and the actual increases appear to lie roughly between the original and revised estimates.

We may consider the relation of the information in Table 1 to that in Figs 2, 3, and 4. We note first that the amplitudes $\sqrt{\alpha_n^2 + \beta_n^2}$ estimated for the longest waves are quite small ($0 \cdot 59$ and $0 \cdot 43$) and that Fig. 4 shows that such values near $2\pi k\xi$ equal to 6 or 7 indicate a directional spectrum approximately equal to $\cos\theta$. Thus from Fig. 2 we would expect the ratio $k\xi/f\tau$, i.e. the ratio of apparent to ideal phase velocities to be about $1 \cdot 15$. This does not agree with the hypothesis at the end of Part I and re-stated in the last paragraph concerning these long waves, but the quite low amplitudes probably result from scatter in the values of R_n and D_n in (8) and (13). The mean value of all amplitudes in Table 1 is $0 \cdot 80$ with standard deviation $0 \cdot 28$, hence the data in Fig. 4 between $2\pi k\xi$ equal to 6 and 16 indicate that the directional spectrum is unlikely to decrease with increasing values of θ less rapidly than $\cos\theta$, i.e. the spectrum is narrower than $\cos\theta$ (and perhaps than $\cos^2\theta$).

For the components near frequency $0{\cdot}78\,\text{sec}^{-1}$, $2\pi k\xi$ is nearly 9, thus from Fig. 2 the velocity increments would be 8 or 9% for $A^2(\theta)$ similar to $\cos^2\theta$ or $\cos\theta$. The observed apparent velocity increment may be expressed as 15 ($\pm$ 5)% roughly. A small increment of only 2 or 3% due to drift velocities, etc. is sufficient to remove any discrepancy, which in any case is within the 90% probability range. The observed apparent phase velocity increases for slightly higher frequencies (table 1), is within that expected for waves with similar directional properties when the increasing effect of drift velocity is taken into account. The observed greater increases at the higher frequencies, if real, may indicate an increasing effect of the drift velocities or of a broadening directional pattern (due to action of smaller scale wind eddies?) for these shorter waves. At longer fetches where the wave system is more advanced, a much greater part of the energy is contained at lower frequencies, and the classical phase velocity of such waves is high (greater than 470 cm/sec for waves of frequency less than $0{\cdot}33\,\text{sec}^{-1}$) relative to the surface water drift, so this feature will be much less important.

One important fact not previously mentioned is that if the actual velocity exceeds the theoretical phase velocities then the wavelengths calculated from a given frequency using the classical formulae are too small. This consideration does not apply to the effect of the multi-directional nature of a wave system. We write for the elevations due to a given component with wavelength λ velocity c

$$h(x,t) \propto \cos\frac{2\pi}{\lambda}(x-ct)$$

If λ_0 and c_0 are the corresponding classical wavelength and velocity then the frequency f is given by

$$f=\frac{c_0}{\lambda_0}=\frac{c}{\lambda}$$

Hence a real velocity increment of $p\%$ implies that the real waves are $p\%$ longer than ideal waves of the same frequency. But the "ideal phase velocity" for waves of length λ is $\sqrt{g\lambda/2\pi}$, thus if we base our calculation on the observation of wavelength, the apparent percentage increase in phase velocity is $\frac{1}{2}p\%$ (since $\sqrt{1+p/100}\simeq 1+\frac{1}{2}p/100$).

It is also important to notice that, in general, the effects of the features mentioned on the rate of transfer of energy in the wind direction by the wave components, will be somewhat similar to those discussed above, except in the following important case. Wave components travelling at an angle to the wind direction transfer energy in their direction of propagation at half their phase velocity, i.e. at the velocity V_g. The rate at which the energy is transferred across a vertical plane perpendicular to the wind direction is $V_g \cos\theta$, and the total rate of energy transfer at a given frequency is proportional to

$$\int_{-\phi}^{\phi} A^2(\theta) V_g \cos\theta . d\theta.$$

Thus the rate of transfer of energy is less than the classical group velocity.

ACKNOWLEDGEMENTS

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The author acknowledges assistance from many members of the National Institute with much pleasure; in particular is grateful to Mr C. M. Hubbard for his assistance in making the observations, and Messrs H. Charnock and M. J. Tucker for their many helpful suggestions. The work in Part 2 has been largely done while a member of the New Zealand Oceanographic Institute, and assistance from other members during the preparation of this paper has been most helpful.

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TYPICAL RAINFALL PATTERNS OVER CENTRAL OTAGO

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Summary

A table has been prepared of synoptic situations associated with rainfalls of 0·25 inches or more at any one of seven selected stations throughout Central Otago. These situation types are discussed with the aid of rainfall charts prepared from observations at approximately 170 rainfall stations in Otago and Southland.

Introduction

Although Central Otago is the driest part of New Zealand, a number of moderate or heavy falls of rain are recorded each year in one part or other of the area. It was the object of this study to classify the synoptic situations giving rise to them and to compare the rainfall distribution occurring in each situation type.

Method

Tabulated daily rainfalls covering a period of 30 months from October 1954 to March 1957 for approximately 170 stations in Otago and Southland were used. Seven stations throughout Central Otago were selected, and every 24 hr fall of 0·25 inches or more, measured at 9 a.m., at any one of these stations, was classified according to the synoptic situation with which it was associated.

A series of rainfall charts was also prepared. When rainfall over any substantial part of Central Otago exceeded 0·25 inches and the synoptic situation was of a simple form, a rainfall chart was drawn, but not when the rainfall was the result of a complex situation involving more than one synoptic feature. Most of the situations occurring on Sundays were abandoned because the data readily available to the writer were incomplete. Some 54 rainfall charts were drawn and these took into account some 60% of the observations of 0·25 inches or more at the selected stations in Central Otago proper, that is, those stations shown in the table below, with the exception of Moa Flat and Wanaka, which are generally regarded as being on the limits of the area. A number of these rainfall charts are reproduced to illustrate typical patterns associated with common situations.

The stations, from which the data were used when classifying the synoptic situations, were Wanaka in the extreme north-west, Cromwell, Alexandra, and Roxburgh on the Clutha River, Moa Flat in the hills south of Roxburgh, and Becks and Ranfurly in the east. The heights of the stations are shown in Table 1.

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TABLE 1.—Number of Days from October 1954 to March 1957 with Rainfall Reaching 0·25 inch and 1·00 inch (in parentheses) in Different Synoptic Situations.

SYNOPTIC SITUATION	MOA FLAT (1,345 ft)	ROXBURGH (350 ft)	ALEXANDRA (520 ft)	CROMWELL (720 ft)	BECKS (1,280 ft)	RANFURLY (1,390 ft)	WANAKA (970 ft)
Cold front from W	27 (3)	19 (2)	12 (1)	16 (2)	19 (2)	19 (1)	40 (4)
Cold front from SW	11	3	1	—	1	1	2
Cold front from S	13 (1)	7	1	1	5	5	—
Air-stream from SW or S	15 (1)	9	—	1	3	2	2
Depression E of S.I.	9 (1)	12	6	6	14	13 (1)	4
Depression crossing S.I.	10 (1)	8	6	9 (1)	10 (1)	10 (1)	7 (1)
Quasi-stat. front	6 (2)	4	4	3	6 (1)	5	6 (1)
Front from NW	—	—	3	6	3	1	10 (2)
Thunderstorms	2	4 (1)	2 (1)	1	2	2	1
Air-stream from NW or W	—	—	—	—	—	—	8
Miscellaneous	2	2	2	—	3	2	—
TOTALS	95 (9)	68 (3)	37 (2)	43 (3)	66 (4)	60 (3)	80 (8)

RESULTS

Table 1 shows the number of occurrences of rainfall of 1 inch or more, as well as the number of falls of 0·25 inches or more. For the purpose of constructing this table, on the comparatively few occasions when the rainfall resulted from a complex situation, an entry was made against the situation type considered to be responsible for most of the rainfall.

The following information on rainfall distribution at the selected stations may be elicited from Table 1:

(1) The greatest number of moderate and heavy falls occurred at Wanaka in the north-west, and at Moa Flat in the south, while there was a steady decline in the number towards the middle of the area.

(2) Cold fronts moving across Otago from the west constituted the most important group of situations, there being a substantial number of moderate or heavy falls at all places, particularly at Wanaka, where the rainfall was largely of orographic origin, in the north-westerly air-stream preceding the frontal passages.

(3) Cold fronts crossing Central Otago from the south-west produced a substantial number of moderate falls only at Moa Flat and Roxburgh in the south, although, when they arrived from a more southerly quarter, a number of moderate falls were also recorded in the east at Becks and Ranfurly. Similarly, most of the heavier falls associated with an air-stream of strong southerly to south-westerly winds were to be found in the southern part of Central Otago.

(4) Depressions east of the South Island accompanied by south-easterly or southerly winds were responsible for a greater number of moderate falls in the southern and eastern districts than in the remaining area.

(5) Fronts, predominantly occlusions extending from depression centres west of Puysegur Point, crossing Otago from the north-west, were responsible for most moderate or heavy falls at Wanaka. The number of these falls decreased steadily towards the south-east and at Roxburgh and Moa Flat no falls of 0·25 inches were recorded.

RAINFALL CHARTS

Some of these situation types will be discussed with the aid of typical rainfall charts associated with them. It may be seen from Table 1 that, apart from at Moa Flat in the extreme south, few falls of 0·25 inches or more were experienced over Central Otago as a result of cold fronts crossing the province from the south-west. Consequently, no examples of this situation are included. In the case of quasi-stationary fronts, the rainfall depended mainly on the location of the front and no useful purpose would appear to be served by demonstrating an example of this situation. Thunderstorms or convective showers were experienced only in summer months, and were not associated with any specific synoptic situation.

However, examples of rainfall distribution associated with the following situations, involving 83% of the observations at the selected stations, are presented.

Cold Fronts Crossing from the West

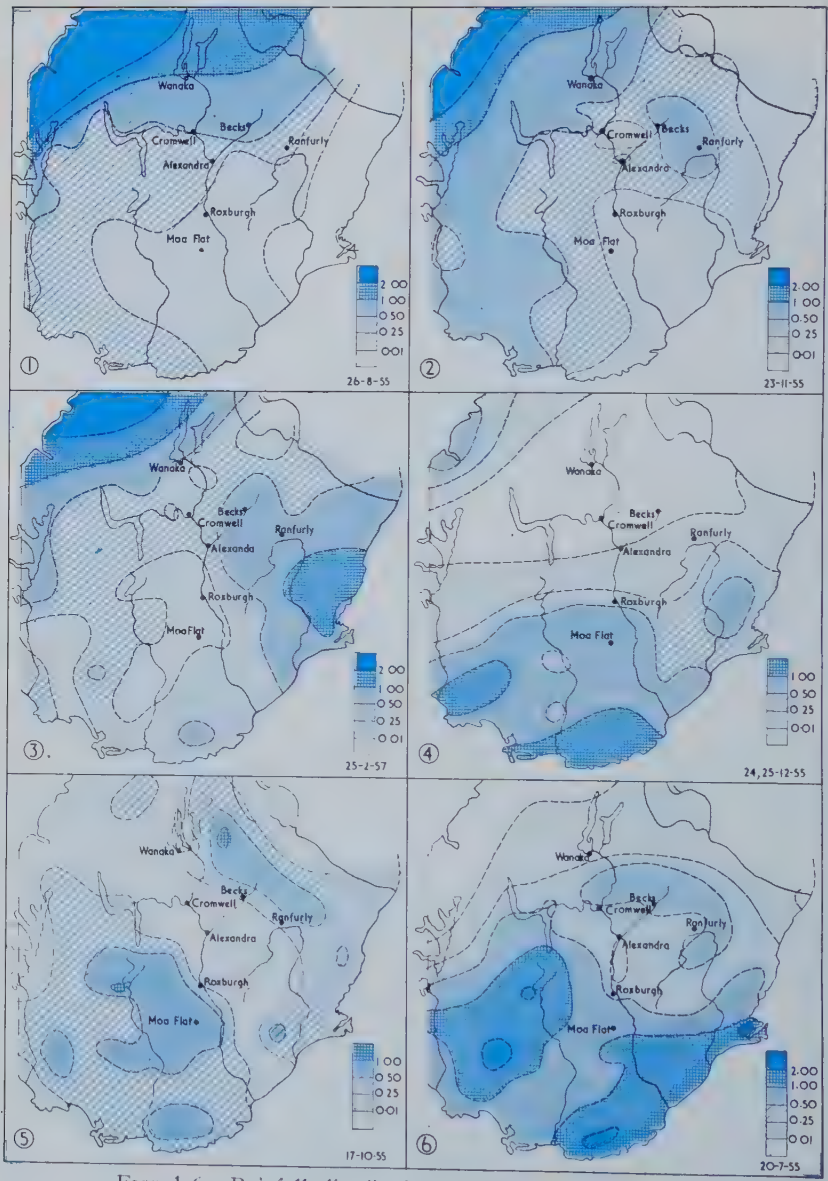
All cold fronts advancing from the west are preceded by a band of rain, heavy in the mountains of the north-west but decreasing towards the east. If the cold front does not suffer retardation during the course of its passage across the South Island, little rain is experienced behind the front except about the south coast. Figure 1 shows the rainfall distribution from a cold front which moved steadily across the South Island, preceded by north-westerly winds and followed by west-south-westerlies. The amount of pre-frontal rain in this example is rather greater than usual. The area of rainfall exceeding 0·25 inches along the south coast of Southland obviously owes part of its origin to post-frontal conditions, as this area is subject to showers in west-south-westerly winds.

When cold fronts from the west are retarded during their passage across the South Island, the rainfall pattern becomes modified by additional rain in the eastern portion of Otago. Two different types of situation arise when a cold front becomes retarded and examples of typical rainfall patterns are illustrated in Fig. 2 and 3.

In the case of Fig. 2, a cold front passed over the province from the west, and after its passage southerly winds developed over the northern half of Otago and over Canterbury, as a shallow frontal wave formed to the east of Canterbury. Superimposed on the normal pre-frontal rain pattern is an area of rainfall exceeding 0·50 inches in the Becks-Ranfurly area, a feature common to all situations of this type.

The other variation of a cold front retarded in its passage from the west is responsible for most of the heavier Central Otago rainfall. In this situation, the cold front moves steadily across Otago, but its progress across South Westland ceases for a time and a small wave depression forms there. That portion of the front over South Canterbury becomes stationary and rain sets in over most of Otago. On 25 February 1957, corresponding to Fig. 3, a closed isobar of 1,000 mb was drawn on the midnight synoptic chart just south of Hokitika. Six hours later this small wave had moved to the east of Canterbury, while the northern portion of the cold front lay oriented north-south through Cook Strait. Note the area of rainfall exceeding 1 inch extending northwards from Dunedin and again the substantial falls in the Becks-Ranfurly region. The amount of rainfall over eastern Otago depends mainly on the amount of progress the front has made up the east coast before it becomes quasi-stationary with the formation of the wave.

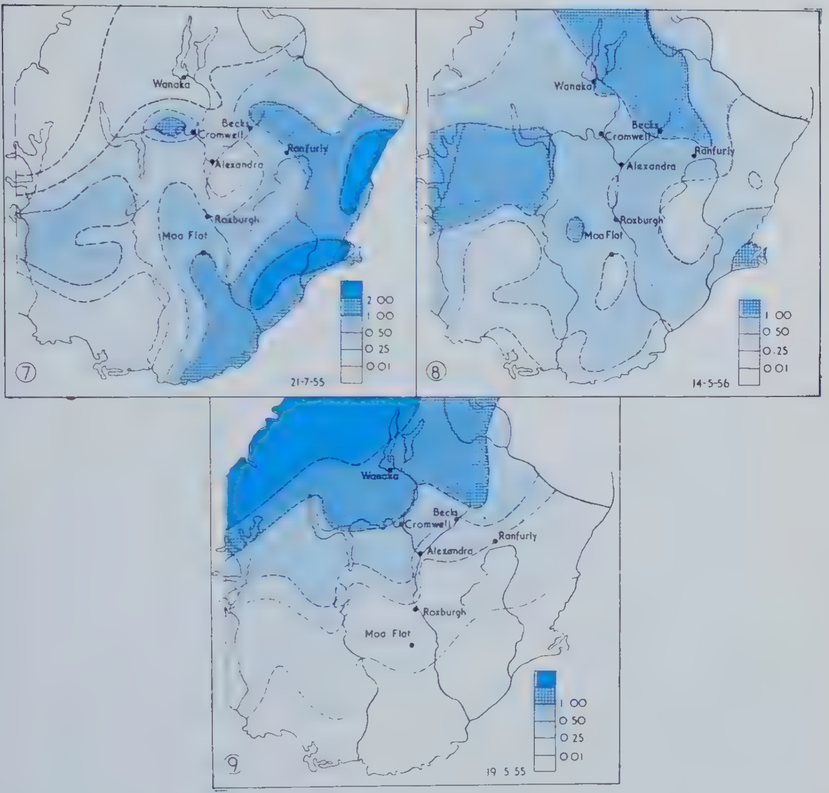
A similar rainfall distribution arises when a cold front moves across Otago but is retarded in the west due to the presence of a depression over the Tasman Sea.



FIGS. 1-6.—Rainfall distribution charts (scale in inches).

Cold Fronts Crossing from the South

Cold fronts advance across Otago from the south when depressions are centred to the east or south-east of the South Island and an anticyclone covers the western Tasman Sea. Figure 4 illustrates the normal



FIGS 7-9.—Rainfall distribution charts (scale in inches).

rainfall distribution in this situation—maximum falls along the south coast and a steady decline in rainfall towards the north, with no rain at all over much of central Otago. In this example the rainfalls for two days have been combined, as the front had barely reached the Waitaki River when observations were taken at 9 a.m. on 25 December 1955. The rainfall north of Dunedin is rather heavier than usual, as the winds behind the front were from the south-south-east.

Southerly or South-westerly Air-stream Showers

Rainfall charts drawn for this situation show great variations throughout the area and comparatively little resemblance from one chart to the next. The only feature the charts have in common is a maximum area extending northwards from a line drawn from Tuatapere to Dunedin. This is illustrated in Fig. 5. The maximum area north of Becks and Ranfurly is not a regular feature of this situation.

Depressions East of the South Island

Figures 6 and 7 represent the rainfall that occurred over Otago and Southland when a depression was centred to the east of the South Island. On 20 July 1955 (see Fig. 6), strong southerly winds gradually backed to south-easterlies and heavy rain fell about south-eastern Otago and over central and western Southland, while there was a decrease in rainfall towards Central Otago. On other occasions when south-easterly winds were less strong, there was less rain over the western part of the chart.

On 21 July winds over the area were strong east-south-easterlies and Fig. 7 shows that the area of heavy rainfall moved northwards up the east coast of Otago, although there is the same decline in rainfall towards Alexandra. On both days substantial falls occurred in the Becks-Ranfurly region of Central Otago.

Depressions Crossing the South Island

The rainfall distribution associated with depressions crossing the South Island depends on the path of the centre. Figure 8 represents the rainfall from a deep depression which crossed from Haast to the east of Timaru and continued to move away towards the south-east. The depression began to cross the South Island on the morning of 14 May 1956. By noon, it had centres of 980 mb east of Timaru and 978 mb just west of Haast. At midnight the depression was centred well to the east of the South Island. The rainfall chart therefore shows the rain that fell from shortly after the depression started to cross the South Island until near the end of the showery period in the southerlies at the rear of it.

The maximum area in the north-west was presumably associated with the western centre of the depression, while the area exceeding 1 inch about Dunedin occurred when south-south-westerly winds developed. The lesser rainfall in the vicinity of Oamaru indicates the degree of sheltering of this area from winds from that quarter.

Fronts Crossing from the North-west

Figure 9 is typical of the rainfall distribution associated with a vigorous occlusion crossing Otago from the north-west, the depression centre lying to the west of Puysegur Point. There is a steady decline in rainfall towards the south-east, and in this case an extensive area over which no rain fell. On some occasions lesser amounts of rainfall are recorded, but the pattern remains similar with the isohyets more crowded in the north-west. At other times, scattered light rain falls in the east and south-east of Otago.

CONCLUSION

Although this investigation is somewhat subjective in character, and rainfall charts were not drawn for every rain-producing synoptic situation, it is considered that the charts illustrate typical rainfall patterns associated with the more prolific of the rain-bearing situations over Central Otago.

PLEISTOCENE AND RECENT STUDIES OF WAITEMATA HARBOUR

PART 2—NORTH SHORE AND SHOAL BAY

By E. J. Searle, Department of Geology, University of Auckland

(Received for publication, 15 July 1958)

Summary

The peninsula of North Shore is a spur of Tertiary sedimentary rocks (Waitemata Formation) that has been subject to considerable marine erosion along its northern side. Deep excavation by an ancient Shoal Bay Stream and its tributaries during glacial times divided the spur into blocks which became isolated by subsequent rise of sea-level. Two groups of volcanoes were active in Recent times and further modified the peninsula.

Introduction

The Takapuna-Devonport peninsula confines the Waitemata Harbour on its northern side. It is a narrow, irregularly shaped landmass of sandstones and mudstones of the mid-Tertiary Waitemata Formation together with accumulations of Pleistocene and Recent sediments, and volcanic materials produced by two groups of eruptive centures. One its northern side, between butresses of resistant volcanic rock at North Head and at the northern end of Takapuna Beach, the soft sedimentary rocks have been worn back to a relatively straight coastline by the waters of Hauraki Gulf. Vigorous marine erosion on this exposed shore has driven the coast back many miles from the original seaward margin of the terrain of Tertiary sediments. In contrast, the south shore of the peninsula, within harbour limits, appears to have retreated relatively little, for although the coast is cliffed for much of its length, wave-cut platforms are not extensive and the offshore profile is steep (Searle, 1958). The shoreline on this harbour side of the peninsula is deeply embayed and the extensive, shallow mud-filled and widely splayed inlets of Shoal Bay and Ngataringa Bay (Fig. 1) extend back into the peninsula almost as far as the northern seaward coast.

The peninsula is, essentially, a long spur of Waitemata sediments which has been partly drowned as a result of post-glacial rise in sea-level. The northern flank of the spur has been removed by coastal recession so that the drainage of the area now almost wholly discharges into Waitemata Harbour. Its southern flank was deeply sculptured by the tributary streams of the ancestral Waitemata River, with the result that large transverse valleys transect the spur and divide it into a number of residuals. The general accordance of ridge levels on these

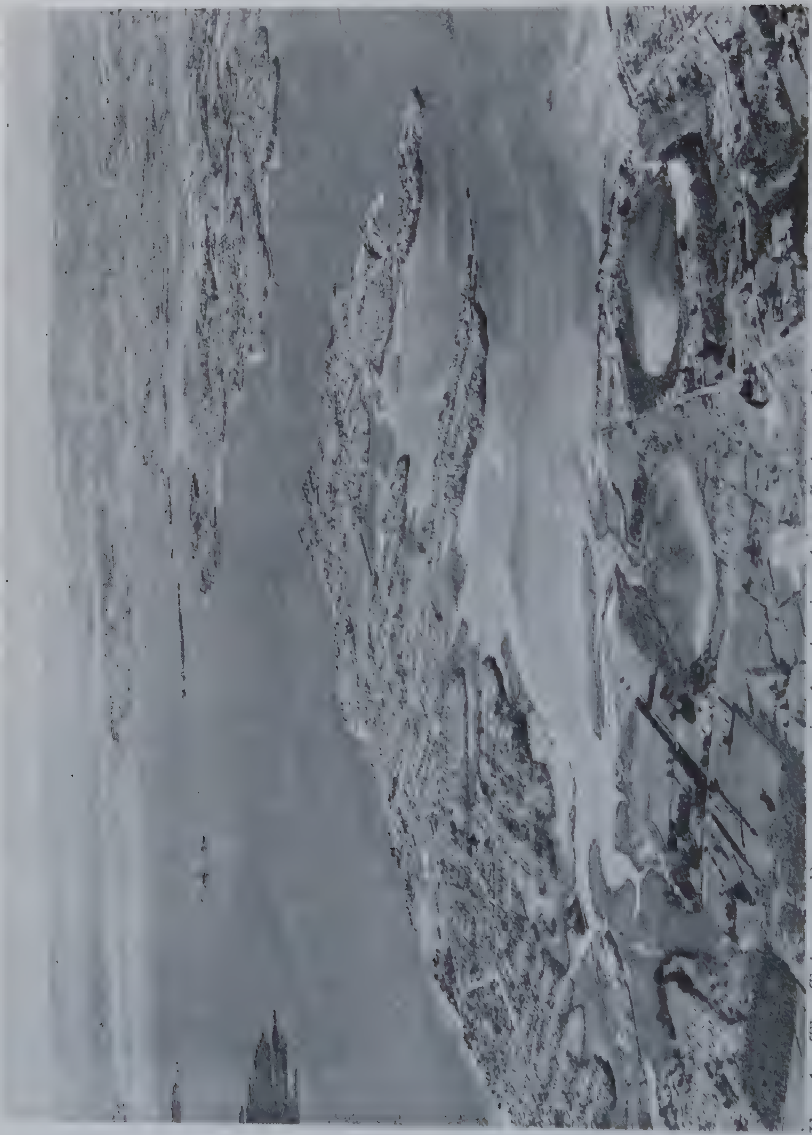


FIG. 1.—The Takapuna-Devonport peninsula, showing the tuff rings of Onepoto (right) and Tank Farm, in the foreground, and Shoal and Ngataringa Bay beyond.
Photo: White's Aviation.

residuals at approximately 100 ft above mean sea-level suggests that the area was part of the 100 ft erosion level so prominently developed in other localities of the Auckland district. The depth to which the tributary valleys were eroded during periods of lowered sea-level in Pleistocene times, together with their considerable width, indicate that they once had much more extensive catchment areas than at present.

The coastal recession that has destroyed their original northern and eastern headwaters has driven the coast back so far that the shoreline now cuts across the valleys where their floors are well below sea-level. Were it not for volcanic debris and other deposits with which these valleys were filled before the present higher sea-level was attained, arms of the sea would now cross the peninsula at several points. The effect of this deep transverse dissection, coupled with the subsequent rise in sea-level, has been to reduce the spur to a series of "islands" of Tertiary sandstone which are tied together by younger sediments and volcanic materials.

The map (Fig. 2) shows four such islands. One of the larger, constituting a large part of Devonport, is connected by a narrow isthmus to

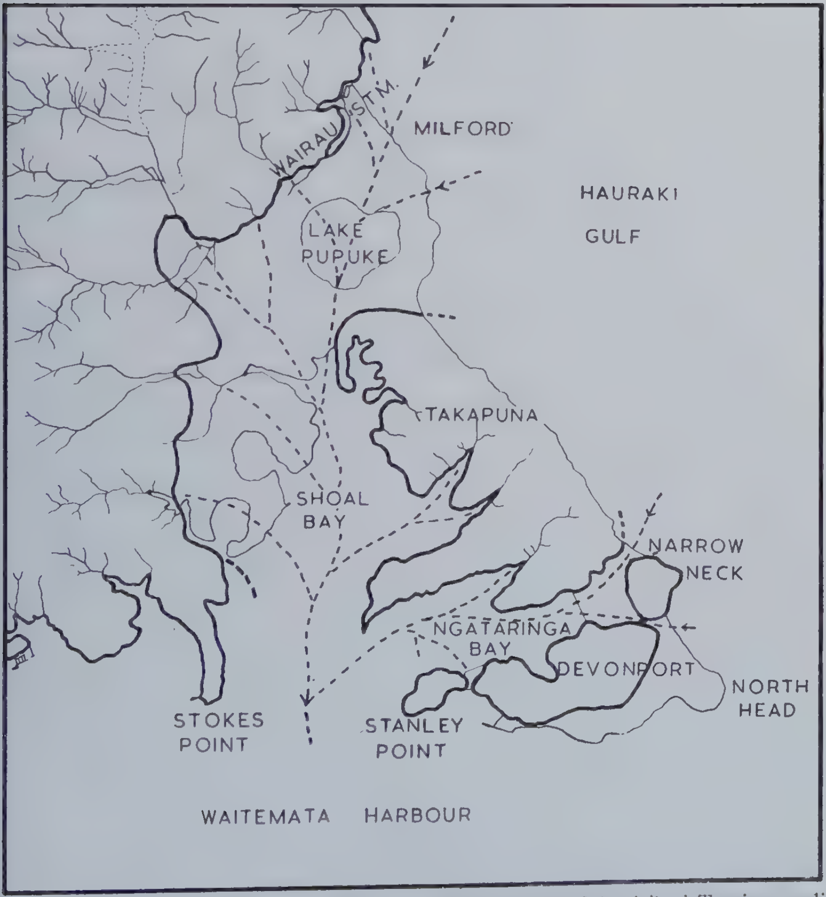


FIG. 2.—Sketch map of Takapuna-Devonport peninsula. "Islands" of Tertiary sediments heavily outlined; present drainage system lightly drawn; probable stream system during low sea-level regime indicated by broken lines.

a smaller island making up the district of Stanley Point. Mr A. T. Griffiths, Engineer to the Borough of Devonport, informs the writer that this isthmus is underlain by marine muds, to a depth of more than 14 ft, so there can be little doubt that a channel between Ngataringa Bay and the main channel of the harbour at one time separated the Stanley Point island from that at Devonport.

A similar low and narrow saddle connects the Devonport island on its northern side with a third block extending between Cheltenham and Narrow Neck. The map of Hochstetter (1864) shows this connection as composed of Quaternary sediments, but these are very thin, for recent drilling on the neck passed through only 16 ft of whitish silts with thin peaty clays at their base before entering Waitemata sandstones at 5 to 6 ft above M.S.L.

Narrow Neck is a sand barrier tying the last mentioned island to the much larger one of Takapuna. The barrier is backed on the harbour side by a lagoonal area much of which, with a surface originally below M.S.L., has been reclaimed; the neck itself rests upon Waitemata sandstone at a depth of approximately 15 ft below M.S.L. Although it would appear that a seaway crossed the peninsula at this point in recent times, a remnant of typical peaty Pleistocene sediments preserved on the southern side of the neck shows that at one time these older sediments filled this ancient valley to heights well above modern sea-level. A period of excavation must have intervened before this minor valley was infilled. In a similar way the Takapuna unit is separated from the Tertiary mainland further to the north by a broad lowland area occupied by Pleistocene silts and Recent or sub-Recent volcanic material. The geological significance of the fill in this valley will be discussed later.

As already noted, volcanic activity has been manifest in two localities on the peninsula. East of Devonport volcanism took the form of scoria cone-building eruptions and effusion of lava, whilst north of Takapuna explosive tuff eruptions were dominant. The volcano of Mt. Victoria erupted on the eastern edge of the Devonport island (Fig. 3) and produced a scoria cone now 261 ft high with a small crater which was breached to the east where a flank emission of lava produced a flow of basalt that cascaded down into the valley of the ancestral Waitemata stream. At the northern foot of this volcano another vent built the broad cone of Mt. Cambria (Mt. Heaphy of Hochstetter, 1864) to a height of only 100 ft, and also produced a minor flow. Another vent opened near the toe of the Mt. Victoria flow and produced a very tiny cone, Duder's Hill, that has long since been destroyed by quarrying. Hochstetter (1864) included Duder's Hill in a sketch of North-Shore and described it as the product of a small independent eruption.

The compound scoria mound of North Head was erupted further to the east. Early activity at this vent was largely explosive and built a compact tuff cone, the material of which is strikingly bedded in thin layers, seldom more than one foot thick. This bedding is regarded as evidence of rapid sequential explosions when coarse material was

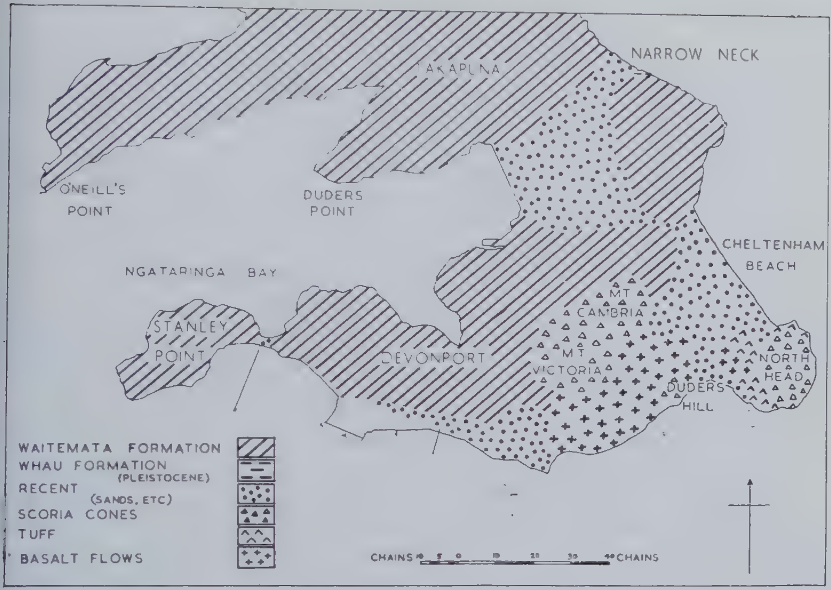


FIG. 3.—Geological map of Devonport.

erupted, interposed with intervals of moderate calm when only fine ash was able to settle. The bulk of fragments in the beds is juvenile in origin and many layers are composed almost exclusively of slaggy basalt and bomb fragments.

This lithology suggests that some fire fountaining accompanied the gas explosions usual in tuff eruptions or, alternatively, that liquid lava was expelled with the gases that blasted the passage to the surface. The pattern of eruption later became more definitely of the normal cone-building type with the result that the earlier tuff cone was partly destroyed by intermittent explosions whilst a steep scoria cone was piled up to envelop much of the older shattered tuff.

The volcanic pile of North Head is separated from Mt. Victoria and its associated vents by a narrow strip of Quaternary sediments. Either the volcanic materials from North Head failed to merge with those from the other volcanoes, or, as is more likely, they made contact below present sea-level. The evidence here, as in most other volcanoes of the Auckland field, is that eruptions occurred at a time when sea-levels were lower than at present and probably even before sea-level had risen sufficiently to enter and submerge the Waitemata Valley. Hochstetter (1864) was misled by the regularity of bedding shown by the tuff of North Head and of other tuff cones in the Auckland district, into assuming that they had been produced by submarine eruptions. He therefore concluded that there had been two phases to local volcanicity, one phase submarine to form tuff cones, and the other supramarine resulting in subaerial scoria cones and lava flows. Such an hypothesis required contemporaneous activity at a large number of submarine centres followed by a very rapid fall in sea-level in order to permit resumption, under subaerial conditions, of activity from the self-same centres in many cases. Such a picture is manifestly unreal and, as suggested above, it is unnecessary to postulate sedimentation through water to account for the regular bedding exhibited by the tuffs.

It seems likely that these volcanic outbursts occurred on the gently sloping northern end of a spur descending into the Waitemata Valley. The eruption of the North Head volcano may well have been the earliest and have taken place well down the spur; later eruptions from the vents of Mt. Victoria and its lesser associates further up the spur, covered the upper slope with the scoria and lava that now mask the southern edge of the peninsula. The rising post-glacial sea-level permitted the accumulation of beach deposits to a height of approximately 10 ft above present mean sea-level where they form distinct terraces. In the area between the Mt. Victoria lavas and the Tertiary sediments of Devonport island a boulder strewn flat was formed where the business area of Devonport is now established. On the northern seaward side of the spur, sands were built up into an extensive terrace between North Head and the Narrow-Neck island along the foot of the cone of Mt. Cambria and the adjacent lava flows. Recent boring here encountered sand only to a depth of 12 ft, indicating that this sand formation extended right across a saddle between North Head and Mt. Victoria. The surface of these terrace formations is accordant with the elevation of another in this area described by Turner and Bartrum (1929) at the south end of Milford beach and of others recently mapped by the writer and Dr R. N. Brothers on the shores of Waitemata Harbour. There can be little doubt that the building of all these terraces is associated with the maximum sea-level at the climax of the Flandrian transgression.

In the northern area of volcanicity (Fig. 4) bordering Shoal Bay and in its northern landward extension, the eruptive pattern was different. Major activity was centred on the vents that formed Lake Pupuke, a sub-circular maar with a saucer-shaped, mud-covered bottom at 200 ft below M.S.L. The margin of this fresh water lake is steeply cliffed and has obviously been enlarged considerably by erosion of the crater walls by the lake waters. Nevertheless, the shape of the perimeter is still sufficiently preserved to suggest that the crater was formed by eruptions from at least two and probably three centres of eruption. The tuff they produced is spread very widely, but there are two features that show that events at this centre were more complex than is generally the case with local tuff-producing volcanoes. In the first place, the fragmental material on the south side of the lake is massive lapilli tuff almost wholly devoid of ash and of accidental rock. Secondly, on all sides the tuff is underlain by flow basalts which are exposed in numerous small outcrops. It seems impossible to escape the conclusion that

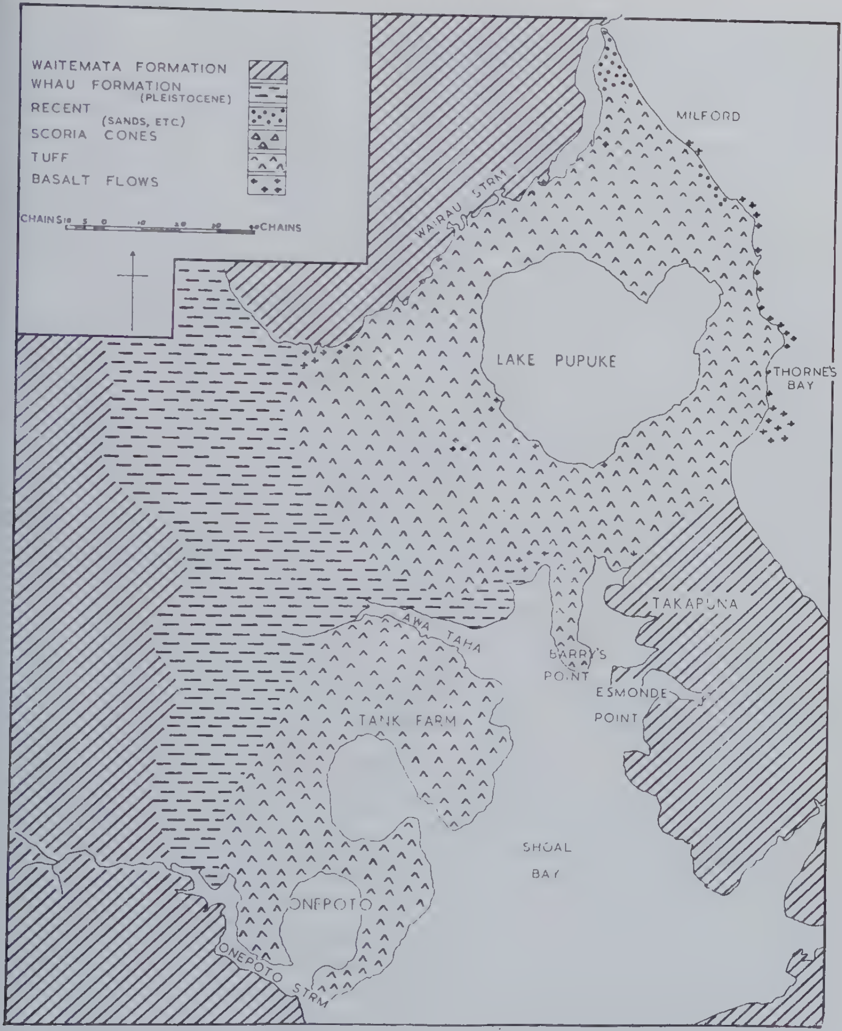


FIG. 4.—Geological map of Takapuna district.

the extensive destructive eruptions that produced both the crater and the tuff had been preceded by quiet outpouring of lava, which spread in thin sheets about the centre, and that accompanying minor explosive activity piled up accumulations of lapilli. The fresh water lake now filling the crater of Pupuke discharges through shrinkage cracks in a lava flow on the shore platform of Thornes Bay, and this indicates continuity of these lavas from crater to coast under the tuffs.

Two smaller explosive eruptions built the low tuff cones associated with Onepoto Basin and Tank Farm and formed craters now breached and occupied by tidal mudflats. That these were simple gas-tension explosions on a relatively small scale is shown by the tuff which on the whole is unbedded, contains relatively little juvenile volcanic material, much fine ash, and a large proportion of accidental rubble.

The inlets of Shoal Bay and the adjoining Ngataringa Bay together cover nearly three square miles. Their common mouth, approximately a mile wide, lies between Stokes Point on the west and Stanley Point on the east, the bays being separated by the mile long peninsula of O'Neils Point. Inside the lines joining these three points practically none of the surface of the bay is below Admiralty Datum level (5·75 ft below M.S.L.) and almost the whole area is exposed at low tide.

Ngataringa Bay is tidal as far as the main north road built across the most easterly bayhead, but its valley continues northwards through the Tertiary sediments and cuts across the peninsula to the opposite coast at Narrow Neck. A side valley makes eastward to Cheltenham and in this fashion the Narrow Neck island has been separated from the other Tertiary blocks of Devonport and Takapuna. Much of the valley area east of the north road was originally a swampy lagoon covered by 5 ft of water at high tide but this was reclaimed many years ago to form a race track now used as a golf course.

Prickings and drilling by Auckland Harbour Board and other authorities indicate that the valleys now occupied by Ngataringa Bay and Shoal Bay have been excavated deeply below present sea-level and later filled with Pleistocene and modern sediments. Midway between Stokes Point and Stanley Point the base of the fill was found at 109 ft below M.S.L., showing that the junction of this valley system was accordant with that of the main Waitemata Valley. Across the mouth of Ngataringa Bay the floor of the valley was at 96 ft below M.S.L. with a small Tertiary rise to 26 ft below M.S.L. midway across. Prickings indicate that the main valley was asymmetrical and cut more deeply on the northern side. Off Duder's Point the channel was cut to 66 ft in Tertiary sediments and the valley bottom steepens gradually so that two miles further upstream where it crosses the northern coast of the peninsula at Narrow Neck it is approximately 16 ft below M.S.L.

Recent engineering work in Shoal Bay has provided much information concerning that area. At the head of the bay tuff and basalt flows from the group of volcanoes already described fill a broad lowland area. The volcanic and pyroclastic rocks rest on a basement of white plastic clays and silts with intercalated beds of peat which are exposed at various points, as for example: beneath the basalts at the northern end of Takapuna Beach; along the margin of the Awa Taha Stream, where they also underlie basalt and tuff from Pupuke; and beneath the tuffs of Onepoto and Tank Farm volcanoes. Similar silts of Whau Formation also outcrop in the low-lying basin north of Onepoto and fill the area between the Onepoto-Tank Farm-Pupuke volcanic accumulations and the high Tertiary terrain farther west.

These Pleistocene sediments have been cut by drilling at many points and appear to have an eroded surface with a maximum altitude of approximately 40 to 50 ft above mean sea-level. Logs of kauri and lumps of kauri gum recovered from beneath the lava flows and tuff deposits, and carbonized logs from the tuff of Pupuke indicate that prior to the local outbursts of volcanic activity, this lowland area was heavily forested. At the north end of Takapuna beach moulds of trunks up to 5 ft in diameter exposed near the foreshore show that lavas eneveloped standing trunks. These lavas overlie thin tuffs at a little below high water level and the trees were most probably rooted in silts beneath the tuffs. This fossil forest is clear evidence that the volcanic activity at Pupuke and the emplacement of the volcanic rocks occurred before the sea attained its present level. Eruptions at Onepoto also overwhelmed standing kauri trees. Recent excavation in the Onepoto tuff has exposed moulds of tree trunks up to 4 ft across, and 47 ft long, and almost vertical in disposition; one horizontal mould is 10 ft in diameter. The walls of the moulds preserve markings of the external features of trees, and bark, root material, and kauri gum have been recovered from the bottoms of the holes. These moulds were completely bridged and sealed by tuff at least 30 ft deep, suggesting that the tops of the trees had been burnt or blasted off by the eruption before the trunks were buried in the debris.

At Onepoto the tuffs and lavas have covered and preserved a remnant of the 40 ft to 60 ft terrace that is so distinctive in the Point Chevalier area on the other side of the harbour. This level has been dated by Brothers (1954) as Main Monastirian and is considered by him to have developed during standstill following a rise of sea-level after the penultimate glaciation.

The Pleistocene sediments that underlie the terrace surface at 45 ft above M.S.L. are exposed on the north-west flank of Onepoto. The following downward succession is recorded in a roadcutting (N42/264662) on Lake Road, Northcote:

3 to 4 ft Tuff from Onepoto centre
(Terrace surface)
15 ft White silt
3 ft Peat
3 ft Silt
16 ft White silt
3 to 4 ft Clays with wood
1 ft Peat
6 to 7 ft Grey silty mudstone
3 ft Peat

Such sediments are typical of a number of lowlying areas at the head of the harbour, about the Tamaki estuary and in districts south of Manukau Harbour. For these sediments the name Whau Formation is proposed and will be fully defined in a later paper.

These Pleistocene sediments are effectively contained within the valley of the ancestral Shoal Bay Stream and rest upon its old floor. At N42/277676, on the north bank of Awa Taha Stream, boreholes from the land surface at about 8 ft above M.S.L. passed through silts to 70 ft below the surface without encountering the local basement Tertiary strata. Near at hand, in other bores, the basement rock was entered at 66 ft below M.S.L. The overall picture suggested by many bores is a very uneven Waitemata surface, on which Whau Formation rests. The Tertiary foundation of Shoal Bay similarly shows well defined and rather sharply-incised drainage channels.

The surface of the Whau sediments also shows evidence of erosion by stream courses and, in the case of the small inlets on the eastern side of Shoal Bay, which have developed along valleys carrying small steep streams tributary to the original Shoal Bay Stream, erosion seems to have removed any sediments of this age that they might once have contained. Nearer the mouth of Shoal Bay the Pleistocene silts are deeply buried under modern muds.

Thus there is a record of three successive systems of channels for the drainage of the area:

(a) the modern streams across the mud flats are channelled in Recent muds;

(b) the muds rest in channels seated in the Whau silts and presumably eroded during the low sea-level times of the Last Glaciation;

(c) the Whau silts, in turn, are contained in valleys carved in the Waitemata strata during earlier periods of low sea-level. As the writer has pointed out (Searle, 1956) similar Pleistocence silts in the New Lynn area, and also at Glendowie, occur at altitudes of over 100 ft above M.S.L., so that presumably they must have been deposited before the 100 ft erosion level was developed. On Brothers' (1954) chronology this would demand the cutting of the valleys in the Tertiary rocks during the low sea-level of the Antepenultimate glaciation at the latest and the deposition of the Whau slits during the succeeding transgression. The sections in Fig. 5 show the relationship of these deposits and valley forms as they are revealed by borings in the bay.

The ancestral Shoal Bay Stream of Antepenultimate glacial times must have eroded a very wide valley extending far to the north-east beyond Milford beach, for its valley there is still very wide at present sea-level. That it was not a seaway and that flow was south into the ancestral-Waitemata River is indicated by the rising Waitemata sandstone floor and by the pattern of drainage to the north of the area (Fig. 1). That the rise in sea-level accompanying subsequent glacial decay was slow is indicated by the silts of the Whau Formation with their numerous peat beds showing frequent swamp conditions. The high still-stand at approximately 110 to 130 ft above modern M.S.L.

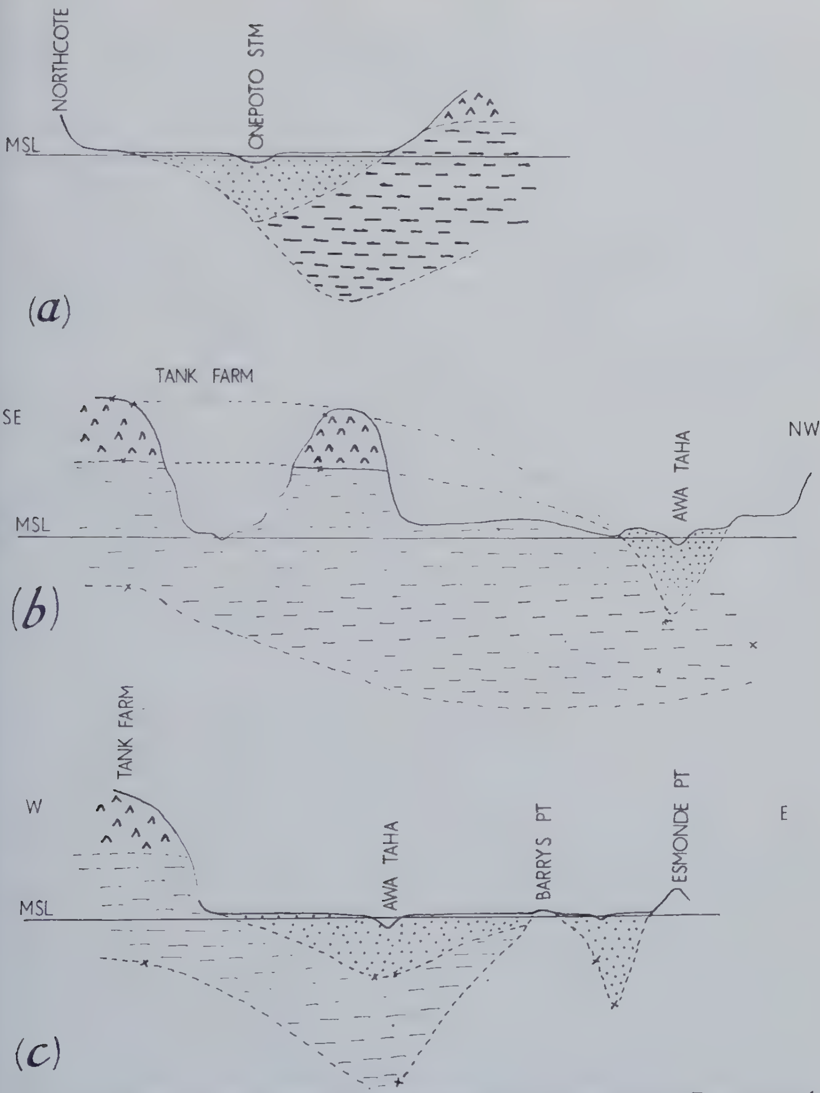


FIG. 5.—Sections across channels discharging into Shoal Bay—Recent muds (dotted); Whau silts (broken lines); tuffs (arrowheads); Waitemata basement (unshaded).
(a) Onepoto Stream. (b) Awa Taha Stream. (c) Tank Farm to Esmonde Road.

during Tyrrhenian times permitted the bevelling of the surface of the Tertiary rocks of the Takapuna-Devonport peninsula, as on those of Auckland isthmus, and is made apparent by the accordance of summit levels on the island-blocks.

During the penultimate glaciation valleys probably were in part re-excavated, but there is no apparent evidence of this erosional interlude yet recognized in the Auckland district. The subsequent period of still-stand at a height of 45 ft to 70 ft above modern sea-level permitted the cutting of a terrace surface at this altitude on the soft Whau silts, as on similar rocks on the southern shore of Waitemata Harbour, but in both areas the harder Tertiary strata escaped extensive planation.

During the last glaciation Shoal Bay Stream and its tributary streams again became incised and cut deeply into the Whau sediments that filled the old valley, so that in the smaller and steeper valleys the Pleistocene silts were completely excavated. The succeeding transgression allowed deposition of muds in these channels and it seems that local volcanoes erupted before rising sea-levels reached their peak.

Turner and Bartrum (1929) and Bartrum (1937) considered the Shoal Bay eruptions to have been the earliest in the Takapuna group, that the white plastic silts (Whau Formation) were deposited about the same time, and that these events preceded the construction of the 40 ft to 60 ft terrace. Later, Bartrum (Clark, 1948) recognized that the evidence on which he had based these views was not reliable and that the eruptions were certainly subsequent to the construction of the terrace surface. The present writer has no doubt that the volcanoes are much the younger features and indeed are probably only a few thousands of years old. The piling up of volcanic products has had a profound effect on the drainage systems of the area north of them. The Onepoto stream was blocked back by the tuff of Onepoto volcano and diverted round its southern margin against the Waitemata sandstones at the base of Stokes Point, to form the present outlet. Erosion and gullying along the flank of the Onepoto tuff ring resulted in the sapping and ultimate breaching of the ring, and its later flooding by tidal water. In a like manner the headwaters of the Awa Taha stream were diverted northwards around the base of the Tank Farm tuff ring which was also breached.

The waters from the head of the Wairau Stream, as pointed out by Turner and Bartrum (1929), originally traversed the area now covered by the basalt flood and the tuffs from the Pupuke vents, and have been diverted northwards around the edge of the Pupuke lava sheet to discharge into Hauraki Gulf at Milford. An alluviated and swampy flat underlain by Whau Formation, and located between the Wairau Stream and those streams discharging into Shoal Bay affords evidence of the earlier course followed by the stream.

ADDENDUM

Since this paper was written, three radiocarbon age determinations of samples collected by the writer in the Takapuna district, and processed by the Division of Nuclear Sciences, D.S.I.R., have recently become available through the courtesy of the Director, New Zealand Geological

Survey. All specimens were obtained from materials associated with the Pupuke volcanics, and whilst they do not precisely date eruptions from that centre they do indicate the minimum age of activity there.

A sample of charcoal (N.Z. ^{14}C No. 227) taken from a lapilli bed in Smale's quarry (N42/277686), gave an age greater than 36,000 years B.P. Wood and kauri gum (N.Z. ^{14}C No. 218), found in peat (N.Z. ^{14}C No. 219), lying immediately below basaltic lava from Pupuke, were obtained from an excavation in the Takapuna Borough Council yard (N42/282679). The wood proved to be older than 42,000 years and the peat more than 40,000 years old.

The weight of evidence would suggest that no great interval of time intervened between the death of the trees and their burial by volcanic products. The peat may possibly be a member of the Whau Formation which constitutes the local sub-volcanic terrain in this locality, but it is more likely to represent superficial carbonaceous debris on the pre-volcanic surface. On the whole it would seem reasonable to accept this data as providing evidence of the minimum age of the lava flows from Pupuke and to deduce, therefore, that the vents which produced the lava were active not less than 42,000 years ago.

Samples of wood and gum have been collected also from beneath the tuff of Onepoto Basin and are, at present, awaiting processing.

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A SHALLOW SHELF AROUND FRANKLIN ISLAND IN THE ROSS SEA, ANTARCTICA

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Summary

A preferred level of marine erosion at a depth of 50 fathoms truncates banks around Franklin Island and forms a shelf similar in form to the continental shelves of temperate regions.

Introduction

In February 1958, H.M.N.Z.S. *Endeavour* made oceanographic and hydrographic observations in the Ross Sea area. While on a reconnaissance of Franklin Island, which lies in Lat. 76° 05′ S, 168° 15′ E, 70 miles east of the coast of Victoria Land and 60 miles north of Ross Island, the ship obtained echo-soundings along a south-south-easterly course while approaching the south-western corner of the island (Figs 1, 2).

Franklin Island was discovered by Sir James Clark Ross in January 1841. He described it (1847) as being 12 miles long (N–S) and 6 miles wide. Its northern shore is a line of precipitous cliffs 500 to 600 ft high, of dark rock with several longitudinal broad white bands and some of a red ochre colour. At the south-western corner, cliffs 400 ft high rise behind a broad level area about half a mile long and some hundreds of yards across, standing about 10 ft above sea-level. Borchgrevink (1901) landed on the island from the *Southern Cross* in 1900 and collected rock specimens. Further collections were made from the *Morning* during Scott's 1901-4 Expedition. Prior (1899, 1902, 1907) determined the rocks as hornblende basalts and a limburgite with fist-sized olivine nodules.

The island is described in the Antarctic Pilot (1948) as being 800 ft high. U.S. Hydrographic Office Chart 6666 gives its dimensions as 6 miles long (N–S) and 3 miles wide.

The island is the eroded remnant of one of the group of basalt volcanoes of probable late-Quaternary age occurring in the western Ross Sea. Photographs indicate that coarse agglomerates, compact jointed basalt, and bedded tuffs crop out roughly horizontally on the face of the steep cliffs at the south-west corner (Fig. 3). The whole island is covered with a thin ice-cap (Fig. 4).

Soundings

Soundings were made on a Hughes recording echo-sounder. The bottom 15 miles north of Franklin Island was in 297 fathoms and along the SSE course was slightly irregular, exhibiting an average micro-

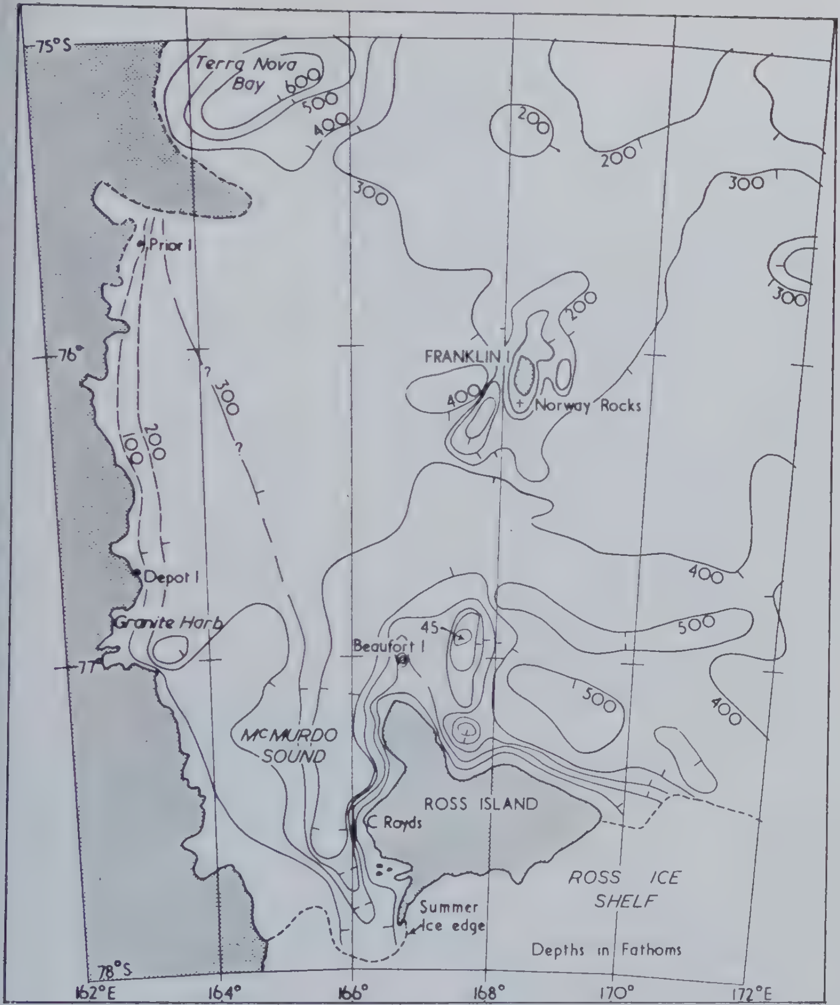

FIG. 1.—Bathymetry of the south-western portion of the Ross Sea and McMurdo Sound. Isobaths at 100-fathom intervals. Data from U.S. Hydrographic Office Chart 6636, Sept. 1957 edit.

relief of 2 fathoms, and having a general concave form. Three miles offshore, in 268 fathoms, the slope (oblique to the north-south trend of the western shore of the island, steepens abruptly; the surface of the slope is highly irregular and it terminates at a break in slope at 50 fathoms 1½ miles from shore. From this point, a smooth shelf slopes gently landward, accidented by minor relief, to 40 fathoms. An increased slope between 40 and 30 fathoms leads to a shallower terrace sloping shoreward (Fig. 5).

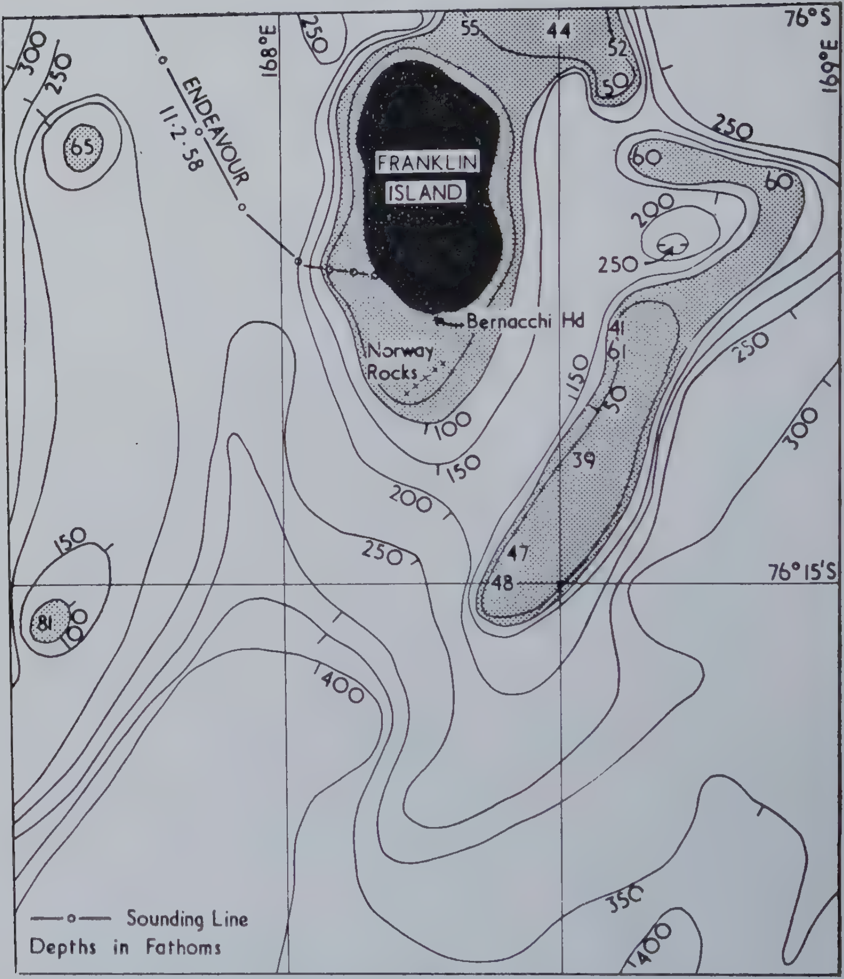

FIG. 2.—The sea-floor around Franklin Island. Isobaths at 50-fathom intervals. Based on soundings on 1955 edition of U.S. Hydrographic Office Chart 6666 and soundings of H.M.N.Z.S. *Endeavour*.

Franklin Island lies roughly at the centre of a 20-mile-diameter group of elongate north-south trending banks that rise to depths of 38 to 81 fathoms from a general sea-floor depth of 250 to 300 fathoms. The island is separated from banks to the west by a 3-mile-wide channel a little less than 250 fathoms deep, and from the nearby south-eastern banks by a channel 200 to 250 fathoms deep and 2 miles wide (Fig. 2). The average depth of the tops of banks, from soundings appearing on the chart, is 52 fathoms.

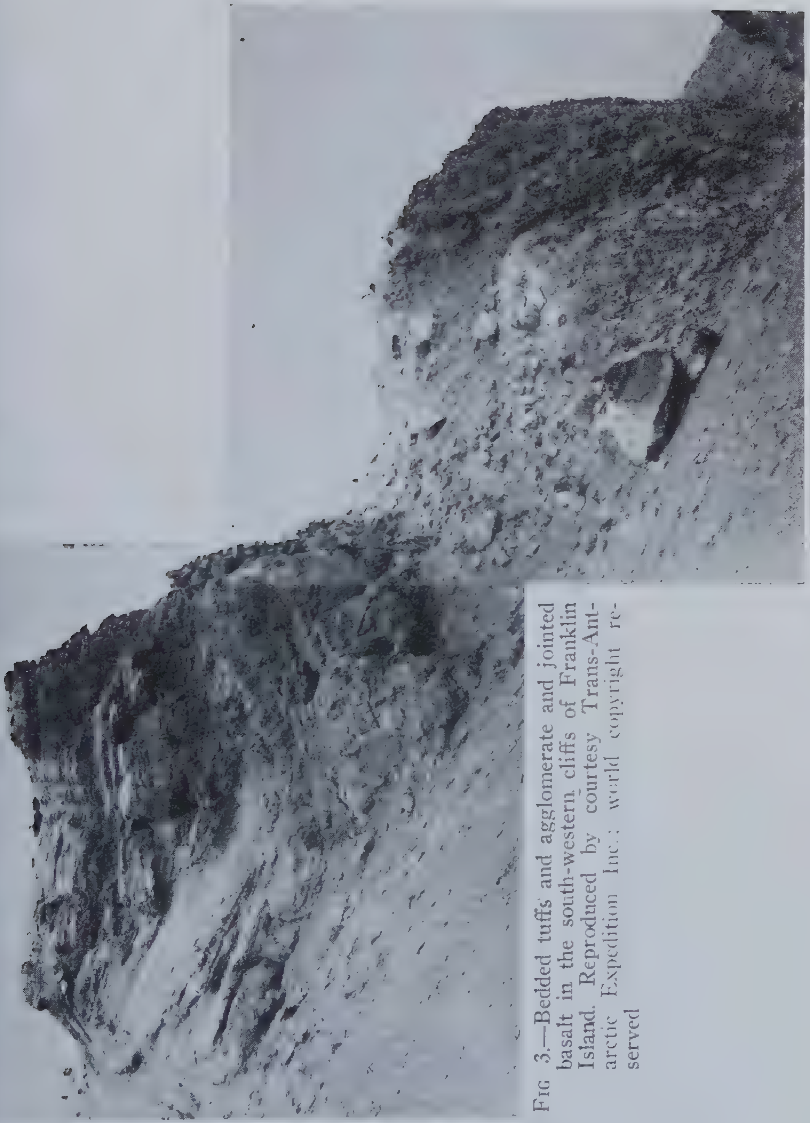

Fig 3.—Bedded tuffs and agglomerate and jointed basalt in the south-western cliffs of Franklin Island. Reproduced by courtesy Trans-Antarctic Expedition Inc.: world copyright reserved

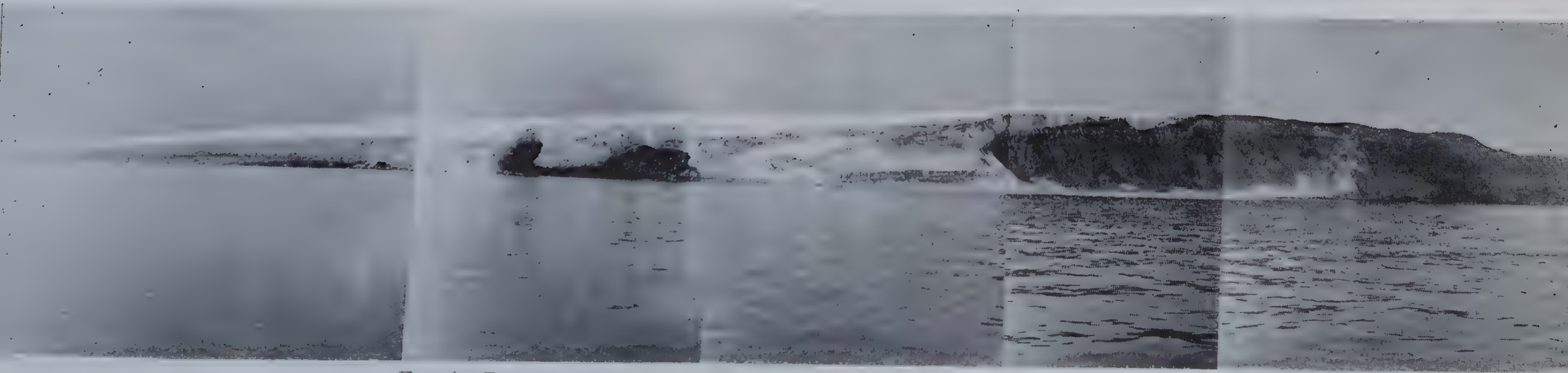

FIG. 4.—Panoramic view of the southern and western sides of Franklin Island. Bernacchi Head is on extreme right. Reproduced by courtesy Trans-Antarctic Expedition Inc.; world copyright reserved.

—*Photograph*: *R. E. Barwick*

DISCUSSION

The group of banks around Franklin Island suggests the broad eroded base of a formerly extensive volcanic pile. The materials of this, as exhibited on Franklin Island, are susceptible to rapid erosion. From the amount of erosion that has occurred, the volcanic mass above sea-level cannot be younger than the Last Glaciation, and is probably of Late Quaternary age.

Morphology

The available data show that the island coast is made up of cliffs, low on the western side (Fig. 4), high on the north and south-west coasts, and (from *Sailing Directions for Antarctica*, p. 197) high on the east coast. At the south-west corner, a phase of talus production has followed the cutting of steep and high cliffs (Fig. 6). The talus apron is now fronted by a flat bench (Fig. 7), itself fronted by low (10 ft) cliffs at the shore-line: the bench gives the impression, sup-

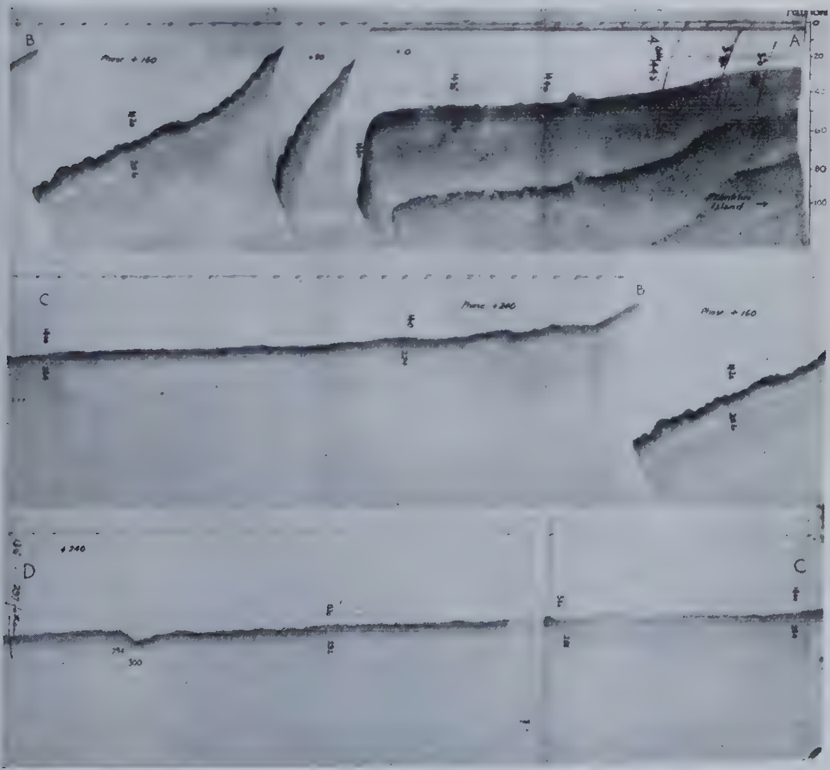

FIG. 5.—Echo soundings obtained from H.M.N.Z.S. *Endeavour* approaching Franklin Island along the course shown on Fig. 2.

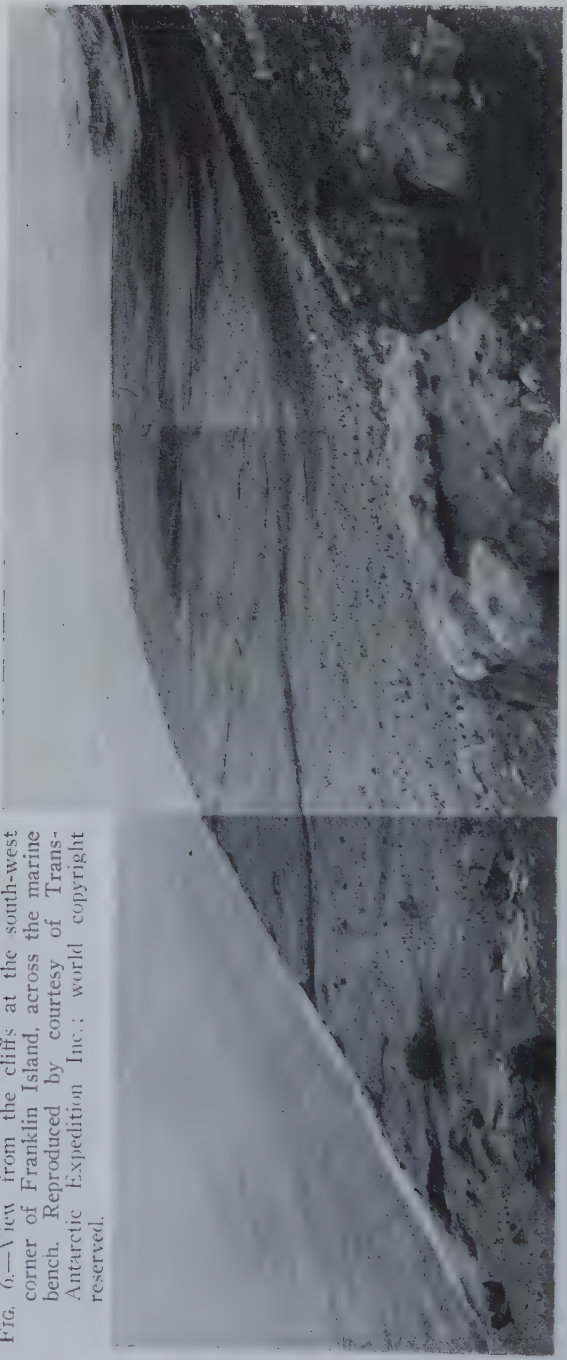

FIG. 6.—View from the cliffs at the south-west corner of Franklin Island, across the marine bench. Reproduced by courtesy of Trans-Antarctic Expedition Inc.; world copyright reserved.

ported by rounded boulders and gravel in its surface, of having been smoothed by marine erosion (Fig. 8). The formation of this raised marine bench can most probably be related to the raised sea-level of the Post-Glacial climatic optimum about 7,000 years ago. It could be analogous to the 25 ft bench described by Harrington (1958) from Beaufort Island. Ferrar (1907) thought the Franklin Island beach similar to that at Cape Adare, which he interpreted as a "re-sorted moraine".

Marine erosion in the volcanic rocks has undoubtedly produced much of the cliffing. The significance of possible ice-erosion cannot be estimated from the data available.

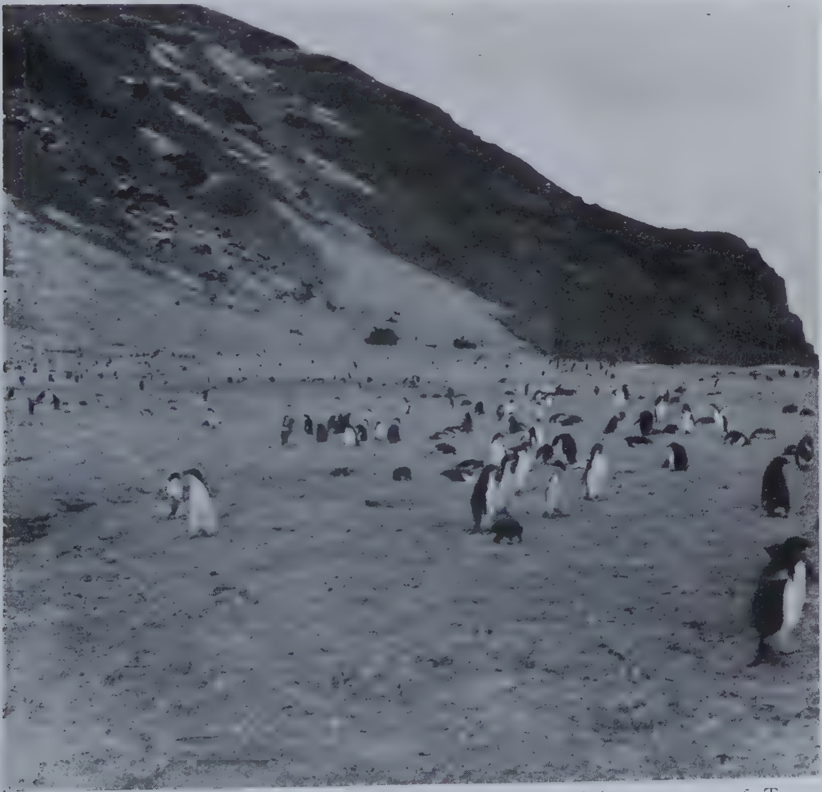

FIG. 7.—Flat surface of the marine bench. Reproduced by courtesy of Trans-Antarctic Expedition Inc.; world copyright reserved.
—*Photograph: R. E. Barwick*

The Deep and Shallow Shelves

Generalizations on the depth of break in slope at the edge of the Antarctic continental terrace have been made by Dietz (1952). Basing his observations on six echo-sounding profiles from the West Pacific and Indian Ocean sectors, Dietz recorded that these showed a break in slope at from 230 to 280 fathoms, that the feature had a wide geographical distribution and of two possible causes preferred to invoke isostatic depression of the Antarctic Continent as a unit, over deep erosion by glacial ice tongues, or by grounded icebergs. The shelf shoreward of this break in slope in the Mackenzie Sea area (Long. 73° E) is flat, and rises gently landward.

This deep break in slope is present in the Ross Sea as a NW–SE trending escarpment dividing the sea floor in 2,000 fathoms and more from the Ross Sea floor of depth 250 fathoms. The break occurs at an average depth of 290 fathoms. The time scale which reaches back to the period of active formation of such a major feature, marginal

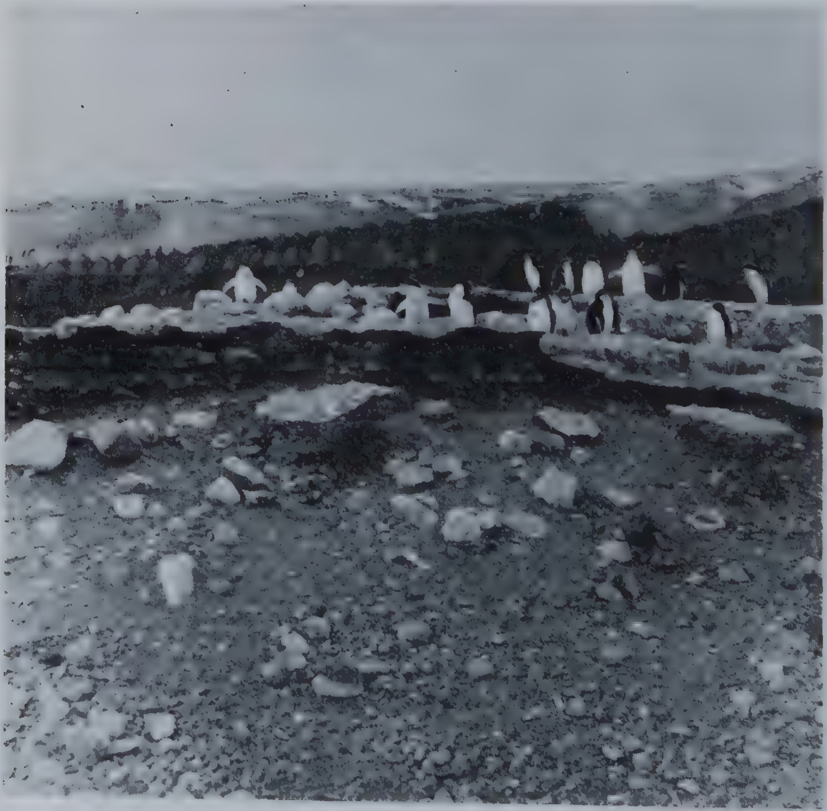

FIG. 8.—Rounded gravels showing imperfect bedding, in sea-cliff at the margin of the marine bench. Reproduced by courtesy Trans-Antarctic Expedition Inc.; world copyright reserved.

—*Photograph*: *R. E. Barwick*

to the whole of the Ross Sea area, and ranging to 500 miles from Ross Island, might well be much greater than that pertaining to fluctuations of ice-cover and morphological changes in the vicinity of the present ice-shelf limits.

The echo-sounding profile and chart soundings show that there is a preferred level developed at a depth roughly of 50 fathoms across the banks near Franklin Island and marginal to the island. The echo-soundings show the shelf surface to be irregular, suggestive of a hard rock surface; the irregularities could also be due to deposition of large ice-rafted blocks.

When a profile across this shelf is compared with a profile of the continental shelf in lower latitudes (Fig. 9), they are seen to be very similar in form. A similar time scale and process could well apply to both features.

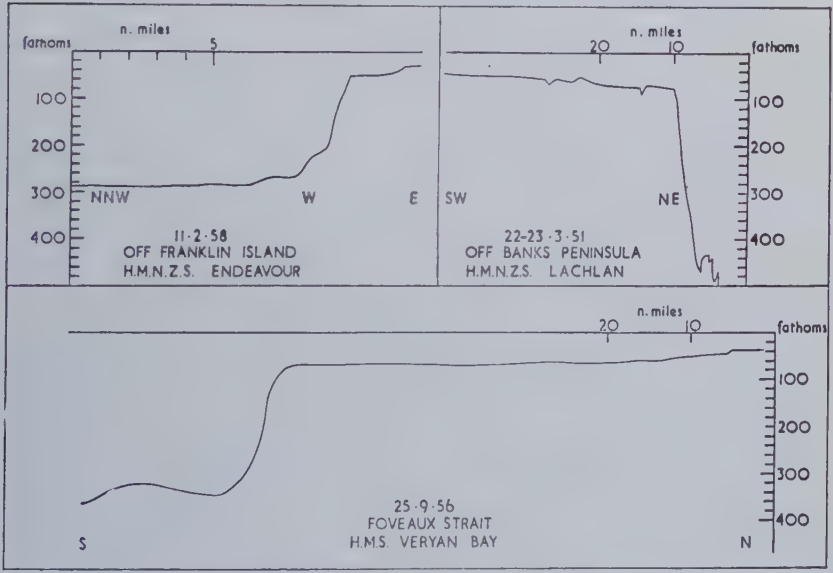

FIG. 9.—Three echo-sounding profiles across shelf margins—Franklin Island, Banks Peninsula, and Foveaux Strait.

Glaciation

The trend of the channels dividing the Franklin Island banks suggests, if these are the direct work of ice, that the direction of movement was to the north. This phase of extension of ice northward can well have been a glacial maximum of the Last Glaciation.

The preferred level at 50 fathoms truncates these banks and channels. While one area alone cannot satisfy requirements for a general hypothesis, it is not unreasonable to postulate that the same processes generally accepted as producing the continental shelf in temperate latitudes have operated also in this region of the Antarctic and at the same period. This implies that when the erosional work of ice during a glacial maximum within the Last Glaciation was brought to a close, the 50-fathom shelf was cut by marine erosion during the rise in sea-level as ice melted. The terrace at 30 fathoms then indicates a pause in the rise of sea-level.

The assumption implies that the area around Franklin Island was substantially free of grounded ice at the time the terraces were formed, and that subsequent glaciation has not been severe enough to modify the shelf then cut.

The average shelf-edge depth for all continental shelves is approximately 70 fathoms. The depth of break in slope here, 50 fathoms, suggests an isostatic uplift in response to a reduction of total ice and water load in this region since the glaciation which preceded shelf-cutting. Alternatively, cutting of the shelf could have been delayed through the protective effect of relict ice until sea-level had risen a little. The same delay mechanism, together with the high cliffs backing the shore and the possibility of reduction of wave action through seasonal ice-cover—all could combine to account for the narrow (1½ mile) width of the shelf.

ACKNOWLEDGEMENTS

The writer wishes to thank the New Zealand Naval Board for the echo-sounding data taken by H.M.N.Z.S. *Endeavour*; Trans-Antarctic Expedition Inc. for permission to publish the photographs of Franklin Island; and Mr R. E. Barwick for his assistance in bringing the photographs to the writer's attention and for his permission to reproduce them.

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GEOLOGICAL INVESTIGATIONS IN SOUTH VICTORIA LAND, ANTARCTICA

PART 1—GEOLOGY OF VICTORIA DRY VALLEY

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(Received for publication, 31 October 1958)

Summary

Large glaciers derived from the ice plateau cut through the mountain ranges of South Victoria Land to the Ross Sea. In the McMurdo Sound region some glaciers have retreated from portions of their courses leaving almost ice free "dry valleys".

The general geology of one such valley, provisionally named here the Victoria Dry Valley, is discussed in detail. The original ice level of this valley was 1,000 ft to 1,200 ft higher than the present remnants of Victoria Glacier. During the summer months meltwater from remnant cirque and glacier ice accumulates as extensive lakes in the depressions of the moraine-strewn valley floor.

The basement is exposed in Victoria Dry Valley as thick uniform sheets of granite, separated by younger dolerite sills. Acidic and basic dykes antedating sill intrusion are associated with the granite basement.

The petrology and field relations of the granite, dykes, and dolerite sills resemble those of similar exposures in the Kukri Hills, 30 miles to the south. It is probable that the rocks of Victoria Dry Valley form part of the northern continuation of the basement complex of South Victoria Land.

INTRODUCTION

General

The McMurdo Sound region of South Victoria Land is an area of mountainous ranges that extend 60 miles inland to merge with the Antarctic plateau at approximately 8,000 ft. Four major glacier valleys cut through these ranges: the Ferrar to the south, the Taylor and the Wright in the centre, and the recently investigated Victoria Glacier* cirque glaciers. The ice free valleys are commonly referred to as "dry valleys" and from air reconnaissance are known to cover some hundreds of square miles in South Victoria Land. During summer months dry valleys are relatively free of snow and ice except for isolated patches

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*New names in this paper have been given by the authors in the field and at date of publication have not been submitted to the New Zealand Honorary Geographic Board.

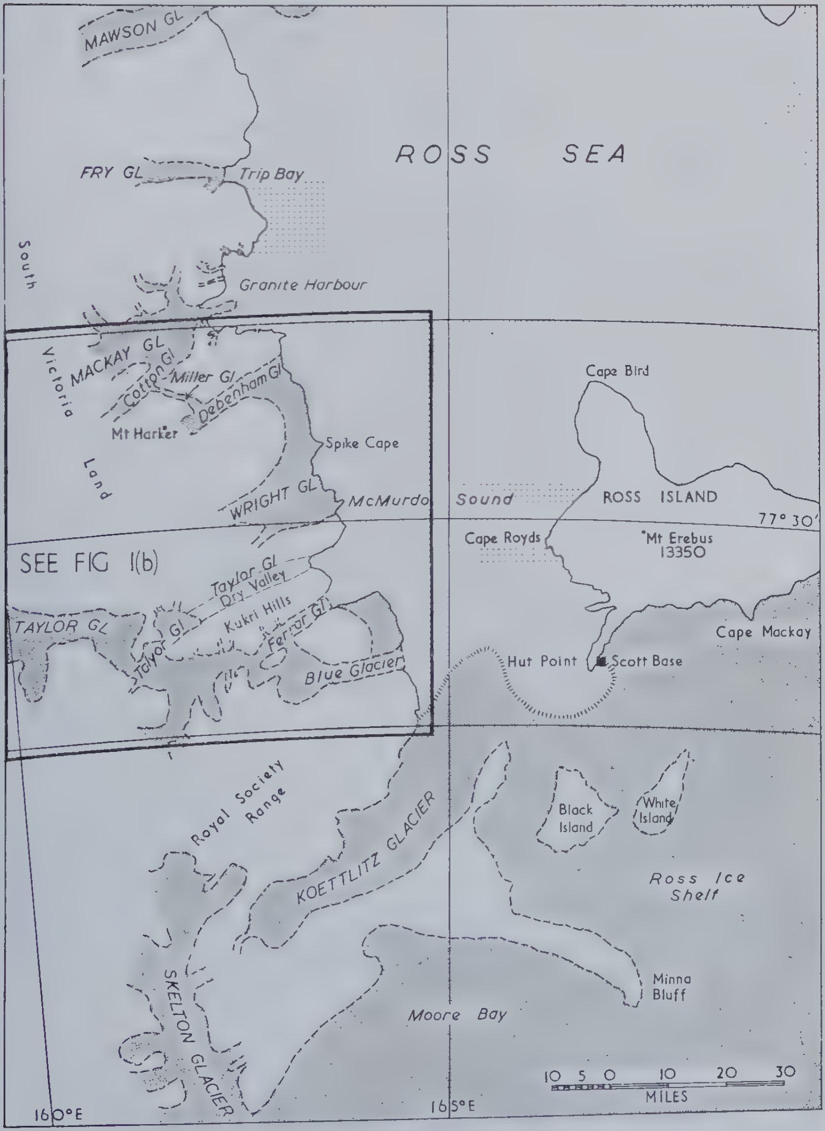

FIG. 1 (a).—Locality map of western coast of McMurdo Sound.

to the north (Fig. 1). Only the Ferrar Valley carries ice throughout its entire length, the others being partly ice free except for hanging within cirques or on glacial shoulders. Lakes commonly occupy depressions in the floors of dry valleys. Shumsky (1957) has described Bunger's oasis, an ice free region in Australian Antarctic territory (lat. 100° 50′, long. 66° 60′) similar to the dry valleys of South Victoria Land.

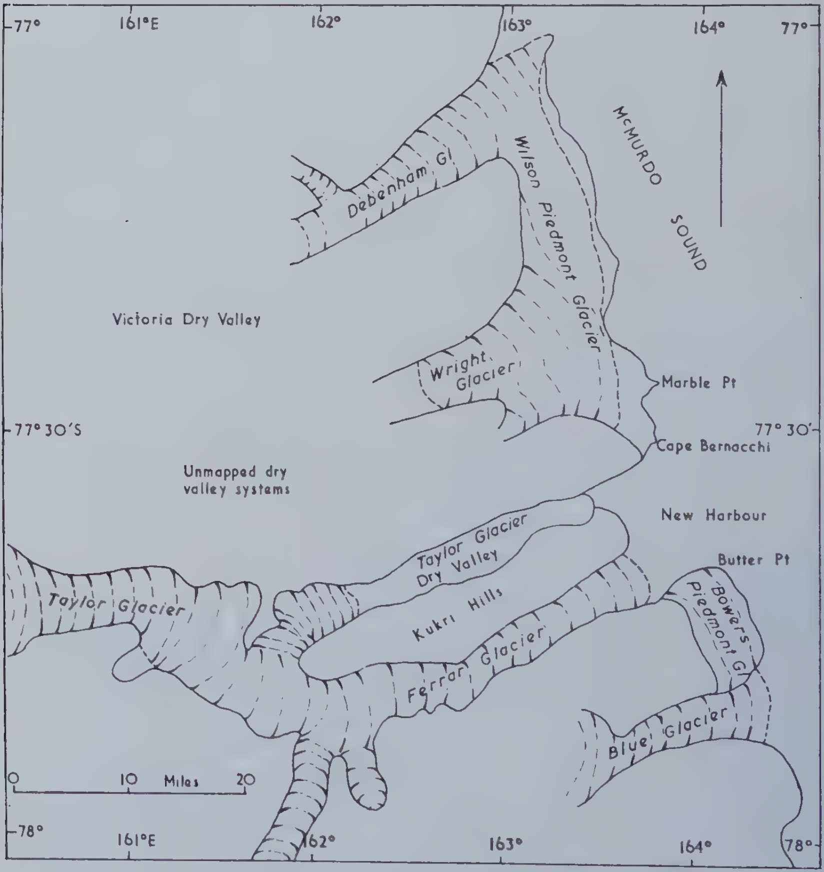

Fig. 1(b).—Victoria Dry Valley area.

Geology of South Victoria Land

An extensive granite and gneiss basement is exposed along the McMurdo Sound coast. Farther inland Beacon Sandstone unconformably overlies the basement (Ferrar, 1907). Extensive dolerite sills, some of great thickness, intrude the basement and Beacon Sandstone.

Location

In January 1958 a base camp was established with the aid of a U.S. Navy helicopter, in a dry valley immediately north of the Wright Glacier at approximately latitude 77° 30′ S and longitude 162° E and at 1,380 ft above sea-level. Victoria Dry Valley* (Figs 1 and 2) and the mountains which rise 4,000 ft above it were examined in the ten days spent at this base.

Previous Work

No geological investigations have previously been made in Victoria Dry Valley. Field parties of Scott's National Antarctic Expeditions of 1901-04 and 1910-14, and Shackelton's British Antarctic Expedition of 1907-09 visited and made a geological reconnaissance of the Ferrar and Taylor Glacier area some 30 miles to the south. This southern area has been visited more recently by geologists of the Trans-Antarctic Expedition.

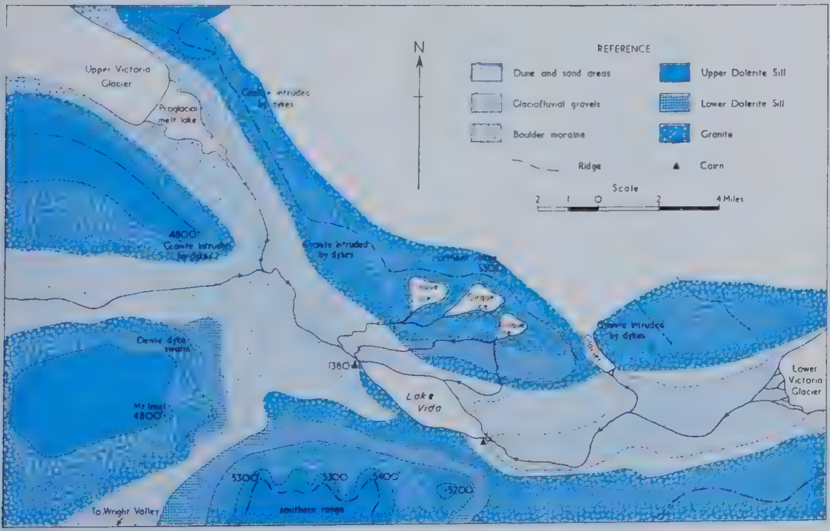

FIG. 2.—Geological map of Victoria Dry Valley.

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*Provisional name.

PHYSIOGRAPHY

The valley of the Wright Glacier extends for approximately 35 miles from the McMurdo Sound coast towards the ice plateau. The entire upper course of this glacier has melted seawards to within 15 miles of the coast. The remnant Wright Glacier is now represented by a stagnant slowly melting mass of ice which coalesces with the Wilson Piedmont Glacier 5 miles from the coast (Fig. 1).

The upper course of the former Wright Glacier is now a wide ice free valley bordered for many miles by high ranges dissected by cirques. Most cirques are ice free by many retain ice tongues, the longer extending down for 1,000 ft into the valley (Fig. 3).

Approximately 20 miles up the Wright Valley from the coast is an

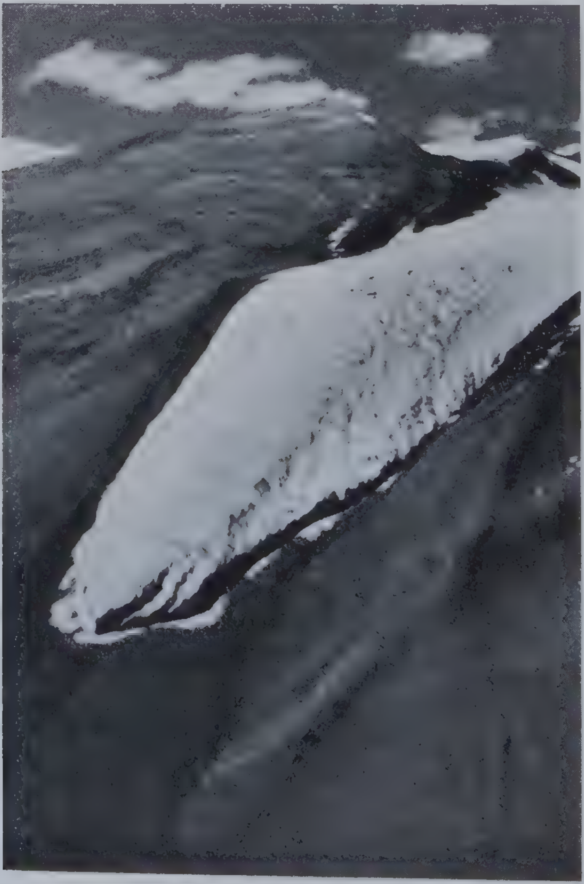

FIG. 3.—Aerial view looking south of an ice tongue descending 1,000 ft down into the Wright Valley.

extensive bifurcating pattern of smaller dry valleys, some of which represent former tributaries of the Wright Glacier. One such valley, the Victoria Dry Valley, was studied geologically in some detail. The glacier formerly occupying this valley and now represented only by its upper and lower remnants is referred to as Victoria Glacier*.

GENERAL GEOLOGY

Dolerite sills are the most conspicuous feature of this region. They are mostly horizontal, range from 200 ft to 1,400 ft in thickness, and alternate with great thicknesses of a coarse homogeneous quartz-orthoclase granite. Acid and basic dykes intrude the granite only.

In addition to the rocks exposed in the valley the moraine contains boulders of porphyritic granite and gneiss with some Beacon Sandstone and it is inferred that these rocks were derived from the interior.

GEOMORPHOLOGY

The Victoria Dry Valley is 6 miles wide and 20 miles long. Westward it rises and branches into three main forks of which the most northerly still contains a glacier, referred to here as the Upper Victoria Glacier* (Fig. 4). This upper remnant appears to be stagnant and retreating although fed by large névé fields only 2 to 3 miles from its terminus. The lower remnant is undoubtedly stagnant and has no large névé fields. It is inferred, but not yet verified, that this portion of the glacier merges with the Wilson Piedmont Glacier only a short distance north of the Wright Glacier. At present the terminal faces of both Victoria glacier remnants are about 20 miles apart and appear to be actively melting back each summer. Meltwater derived from this source drains into Lake Vida* in the centre of the dry valley (Fig. 5).

Valley Walls

Cirques with accordant floors 2,700 ft above the valley floor are conspicuous along the southern side of the valley (Fig. 6). The cirques have gently sloping backwalls and are mostly half a mile long and a quarter of a mile wide. Most cirques contain small lakes, and moraine charged masses of ice lie nearer the backwall.

The shape of the cirque walls is controlled by the rocks in which they are cut. Within granite which is soft and uniform, they are V shaped. When cut through dolerite and granite their cross-sections are U shaped, their upper parts being steeper in dolerite than granite.

A well developed glacial shoulder about 1,200 ft above the valley floor extends along both sides of the valley for most of its length. This suggests a former ice level of 1,000 to 1,200 ft above the present Victoria Glacier (Fig. 7). Within the shoulders small circular depressions have filled with meltwater to form tarns. Other tarns have dried up leaving salt encrusted basins.

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*Provisional names.

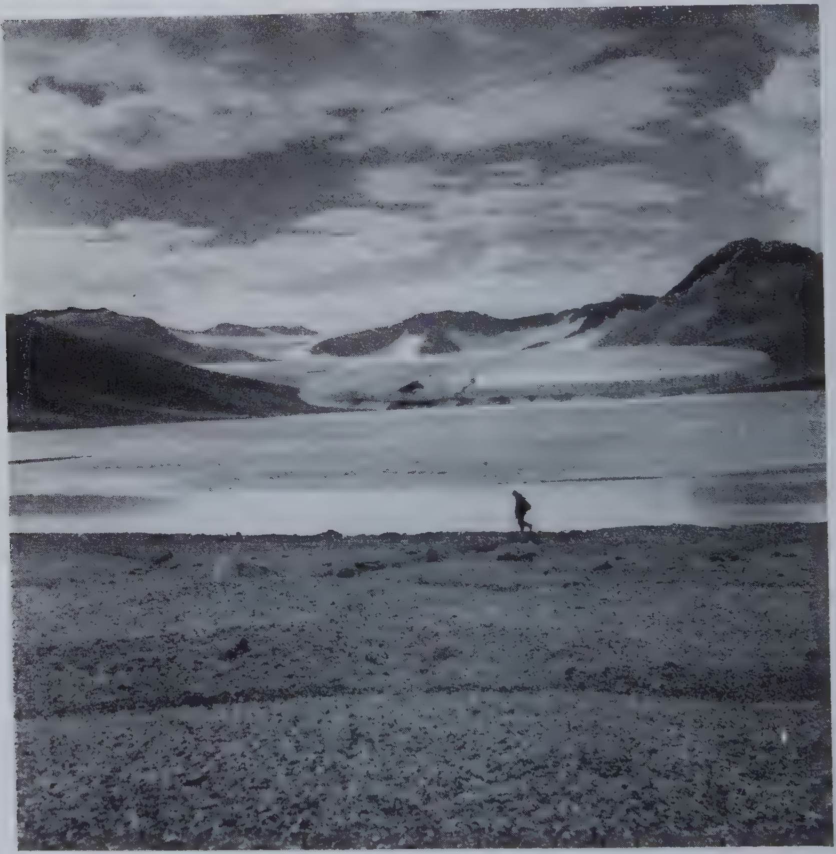

FIG. 4.—View looking north-west at the face of Upper Victoria Glacier. The crevassed terminus spreads laterally over the valley floor, the ice cliffs standing 170 ft high. An ice covered proglacial lake occupies the centre of the photograph; in the foreground moraine is patterned by frost polygons.

The Valley Floor

Undulating moraine extends over some 60 sq. miles of the valley floor and rises in places to 1,000 ft above Lake Vida. The wind polished moraine consists of dolerite boulders up to 3 ft in diameter with a smaller proportion of gneiss and granite erratics. Much of the valley floor moraine and some scree slopes are patterned with frost polygons up to 40 ft in diameter. In profile each polygon is broadly concave and

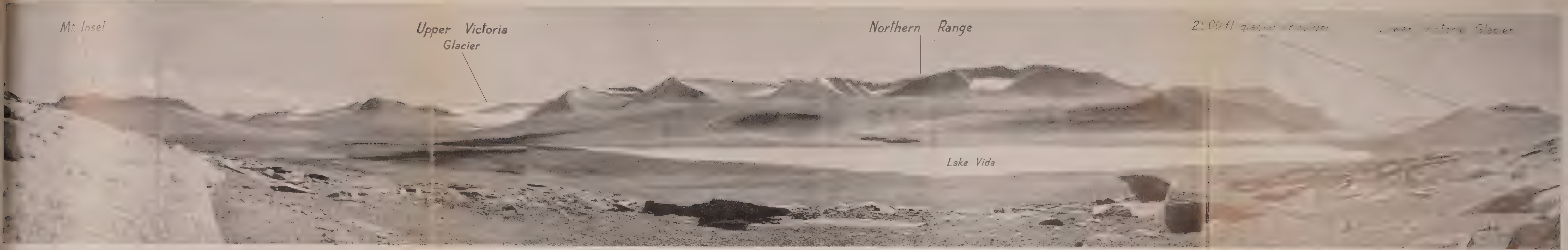

Fig. 5.—View facing north across Victoria Dry Valley showing the upper and lower remnants of Victoria Glacier. Ice covered Lake Vida is fed by melt-water derived from remnant cirque and glacier ice. The glacial shoulder which is present throughout the valley is most evident in the extreme right of the photo where it is 1,000 to 1,200 ft above the remnant Lower Victoria Glacier. The floors of the cirques in the Northern Range are accordant with this well developed glacial shoulder.

Photo: R. E. Barwick

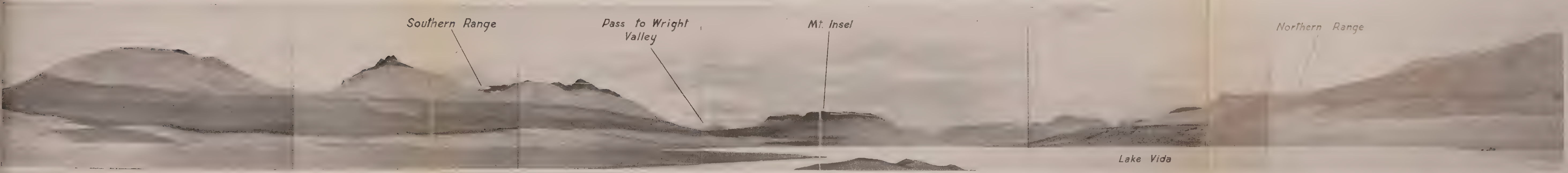

Fig. 6.—View from the eastern end of Lake Vida looking westwards up Victoria Dry Valley. Two dolerite sills are exposed in the walls of the Southern Range, the upper forming a remnant cap above the granite, the lower, about 1,400 ft thick, as a continuous intrusion through the granite. The southern range has been dissected by many cirques which are now almost completely free from ice. In the middle distance, dolerite-capped Mt. Insel forms a divide between the western dry valley forks. To the extreme right a considerable area of sand and boulder moraine has accumulated along the northern shores of Lake Vida.

Photo: R. E. Barwick

separated from adjacent polygons by a sand-filled trough about 2 ft wide. In coarse moraine, polygons are smaller and have deeper peripheral troughs.

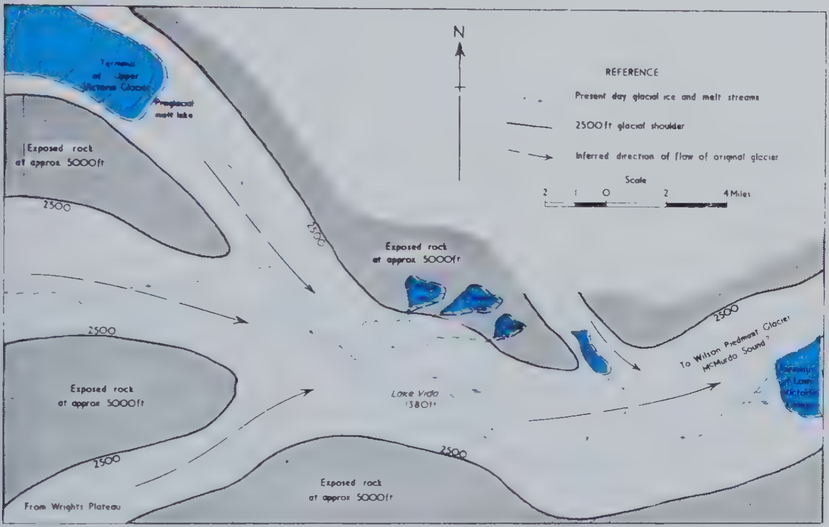
Terminus of Upper Victoria Glacier
Proglacial melt lake
Exposed rock at approx 5000ft
2500
2500
Exposed rock at approx 5000ft
2500
Exposed rock at approx 5000ft
2500
From Wrights Plateau
Exposed rock at approx 5000ft
Lake Vida 1380ft
2500
N
REFERENCE
Present day glacial ice and melt streams
2500ft glacial shoulder
Inferred direction of flow of original glacier
Scale
2 1 0 2 4 Miles
2500
To Wilson Piedmont Glacier McMurdo Sound?
Terminus of Lower Victoria Glacier

FIG. 7.—Shows the extent of a larger glacier which formerly flowed eastwards through Victoria Valley to McMurdo Sound. The present state of this glacier is represented by Upper and Lower remnants of Victoria Glacier.

Sand Areas

During the summer months the permafrost level of the valley floor retreats below the surface. The water-saturated glacial drift is drained by meltwater streams and when completely dry becomes a fine wind-blown sand. To the east of Lake Vida barchan type sand dunes 30 ft to 50 ft high and 200 ft to 300 ft long, associated with longitudinal dunes have been formed under the influence of the local up-valley wind. This entire sand area is underlain by permanently frozen silt. Where sand dunes encroach upon the delta drainage system during the summer flow periods their flanks are quickly cut away.

South-west of the central valley lake near the low pass leading to the Wright Valley from Victoria Valley a wide belt of elongated trough and mound forms are sculptured from moraine debris. The troughs are commonly filled with wind-blown sand that has a surface veneer of ventifacts. The divides are formed of dolerite boulders sandblasted to honeycomb patterns.

REMNANT GLACIAL AND MELTWATER PHENOMENA

At the terminus of Upper Victoria Glacier vertical ice cliffs 100 ft high overlie 70 ft of subglacial moraine which is exposed as a gentle apron sloping down to the valley floor. From Figs 4 and 8 it is evident that the vertical profile of the terminus changes abrupty at the interface between glacier ice and subglacial moraine. It appears that the greater amount of heat absorbed by the subglacial moraine causes undermining of the glacier ice. Vertical jointing of the glacier ice results in the calving of ice slivers from the terminal face on to the apron below. This sudden influx of potential meltwater causes widespread transgressions of the lake shoreline to occur. Streams of meltwater derived from the upper surface of the glacier cascade down the terminal face on to the apron below. These streams flow south-east through a

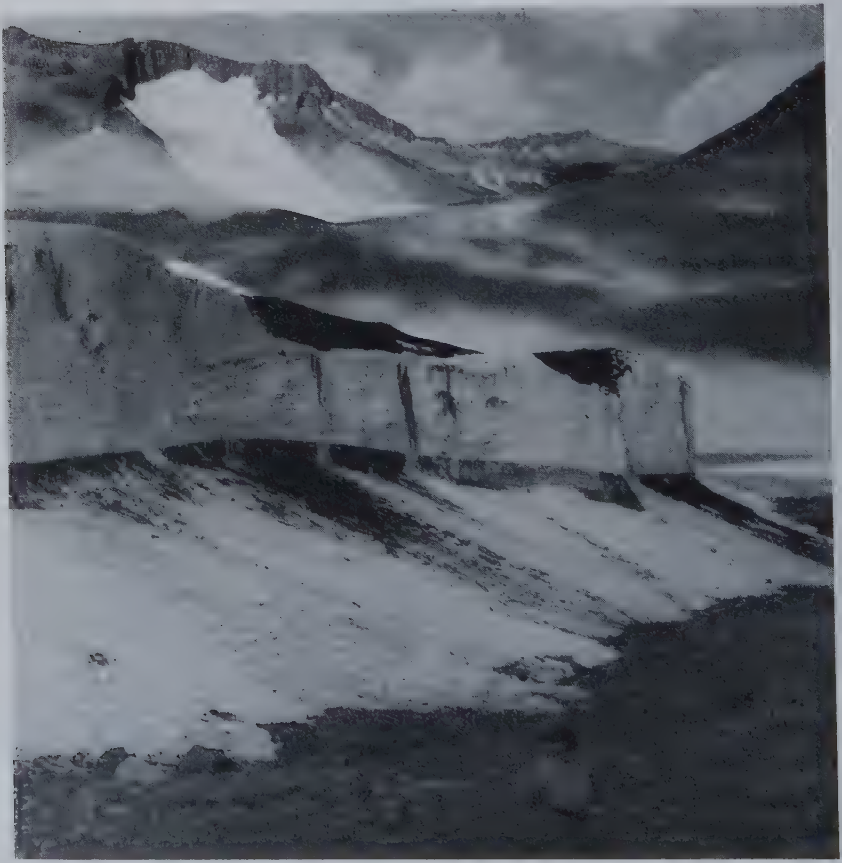

FIG. 8.—The ice cliff terminating the Upper Victoria Glacier rises 100 ft above an apron of subglacial moraine 70 ft in height. Frozen pillars of meltwater descend from the surface of the glacier to the frozen apron of ice blocks and moraine below.

series of small lakes before entering Lake Vida via an extensive silt delta. Meltwater streams are similarly derived from the Lower Victoria Glacier and flow westwards to Lake Vida.

Lake Vida has an area of about 8 square miles, its length (4 miles) being aligned along the valley. Except for a marginal belt of water averaging 20° C the entire lake surface was ice covered in January 1958. During the summer Lake Vida is receiving meltwater (Fig. 9) from cirques along the valley wall in addition to that derived from the upper and lower remnants of Victoria Glacier. The lake shoreline varies from pebbly beaches to frozen silt cliffs 8 ft high. Small nodules of quartz and feldspar grains embedded in a silt matrix are typical of one part of the shoreline (Fig. 10). Small marginal lakes and inlets indicate the existence of a formerly higher lake level.

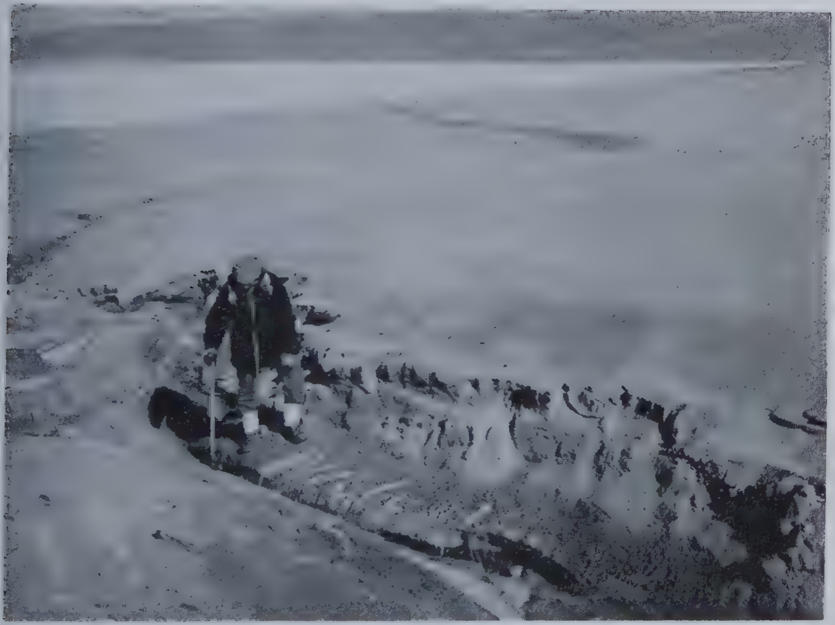

FIG. 9.—A cirque meltstream on the northern valley wall has cut 5 ft into dry wind-blown sand and the permafrost that underlies it. In the middle distance meltwater from this and other meltstreams enters ice covered Lake Vida.
Photo: R. E. Barwick

GRANITE

Two main types of granite were recognized in the McMurdo Sound region by Ferrar (1907). The older, grey in colour, is a quartz orthoclase, oligoclase, biotite granite commonly containing xenoliths of metamorphic rocks. The younger granite consists of hornblende, biotite, orthoclase, oligoclase and quartz, and is believed to form a sheet above the grey granite. In the Kukri Hills (the divide between the Ferrar Glacier and Taylor Dry Valley) the pink and grey granites are separated by a dolerite sill.

A great variety of younger acid and basic veins and dykes including aplites, pegmatite, granite porphyries, porphyrite, and lamprohyre have been described by Prior (1907), Mawson (1916), and Smith (1924).

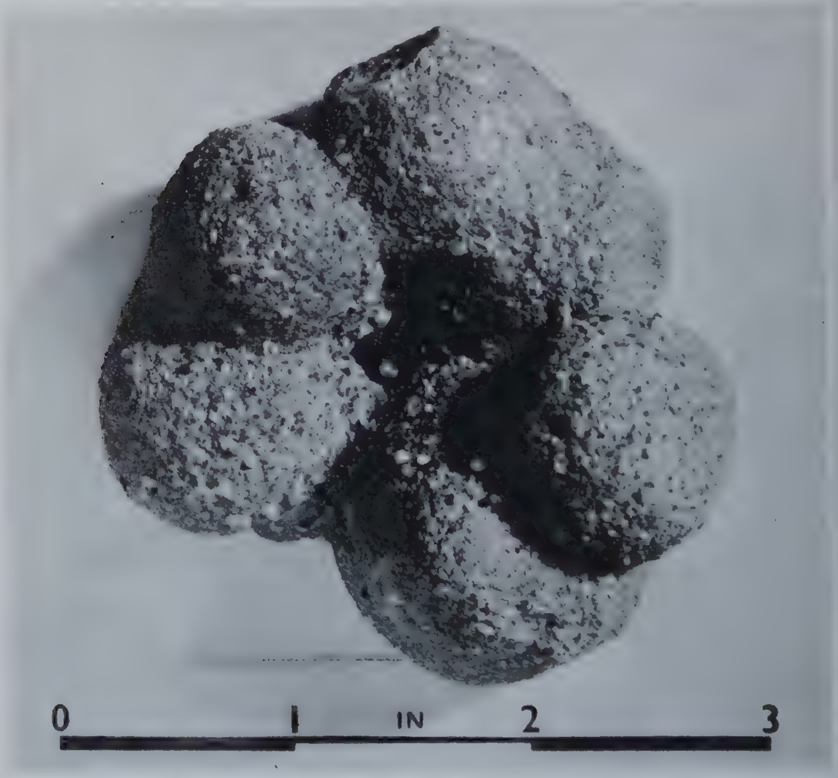

FIG. 10.—Nodules composed of quartz and feldspar grains cemented by clay, occu along the shoreline of Lake Vida.

Photo: M. D. King

Petrography of Victoria Dry Valley Granite

About 2,500 ft of granite is exposed along the southern walls of the valley where it is separated by a dolerite sill into two thick sheets (Fig. 6). The granite at each contact is finer grained and contains a slightly higher percentage of biotite than the typical granite, while the feldspar present has a more intense pink coloration. The granite above and below the dolerite sill is intruded by acid and basic dykes which do not intrude the dolerite, and it is inferred that dykes are older than the dolerite sills.

In hand specimen the typical granite from each sheet is a coarse even-grained rock of white and pink orthoclase and oligoclase with plentiful quartz and scattered shreds of biotite. This rock is mineralogically similar to the grey granite described by Ferrar (1907). A small area of a gneissic rock within the granite is exposed near the western end of Lake Vida. This rock is mottled dark green and white with strongly developed foliation. Mineralogically it consists of equal amounts of oligoclase and hornblende with less abundant biotite. A similar hornblende diorite was described by Mawson (1916) as an inclusion in the grey granite of Depot Island. Ferrar (1907) describes hornblendic inclusions in the pink granite of the Kukri Hills.

Below Mt. Insel, dyke intrusion into granite is particularly concentrated (Fig. 11). Between the parallel basic dykes a laminated granite

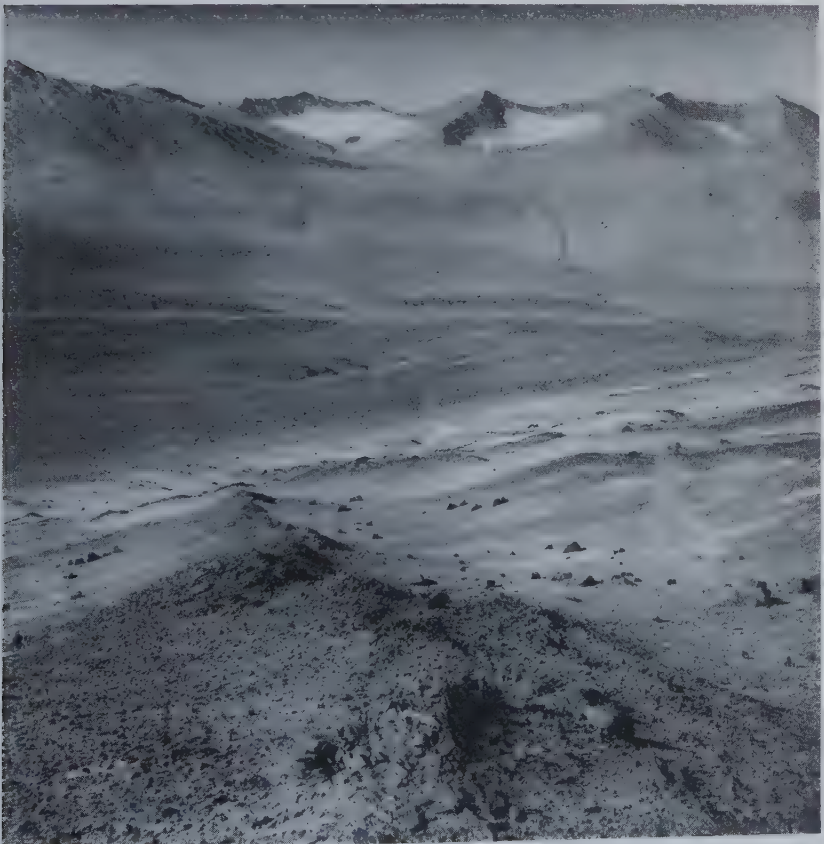

FIG. 11.—View looking north-east across the valley floor to three cirques incised near the top of the northern valley wall. Meltwater courses run from the cirques to the silt plain of the valley floor. The ridges in the foreground are lamprophyre dykes which intrude granite.

consisting predominantly of quartz with lesser amounts of plagioclase, calcite, and biotite is exposed. It is assumed that the intensity of intrusion in the locality has produced this altered rock.

Weathering Forms

The granite exposed in the wall of the Southern Range weathers into heaps of subangular to rounded exfoliation slabs usually stacked convex side uppermost. Within the cirques of this range a mushroom weathering form predominates (Fig. 12). The resistant cap of each mushroom is iron stained and little abraided, while the lower cavernous sides have been actively sandblasted. Freeze and thaw phenomena also play an important part in the disintegration of the friable granite.

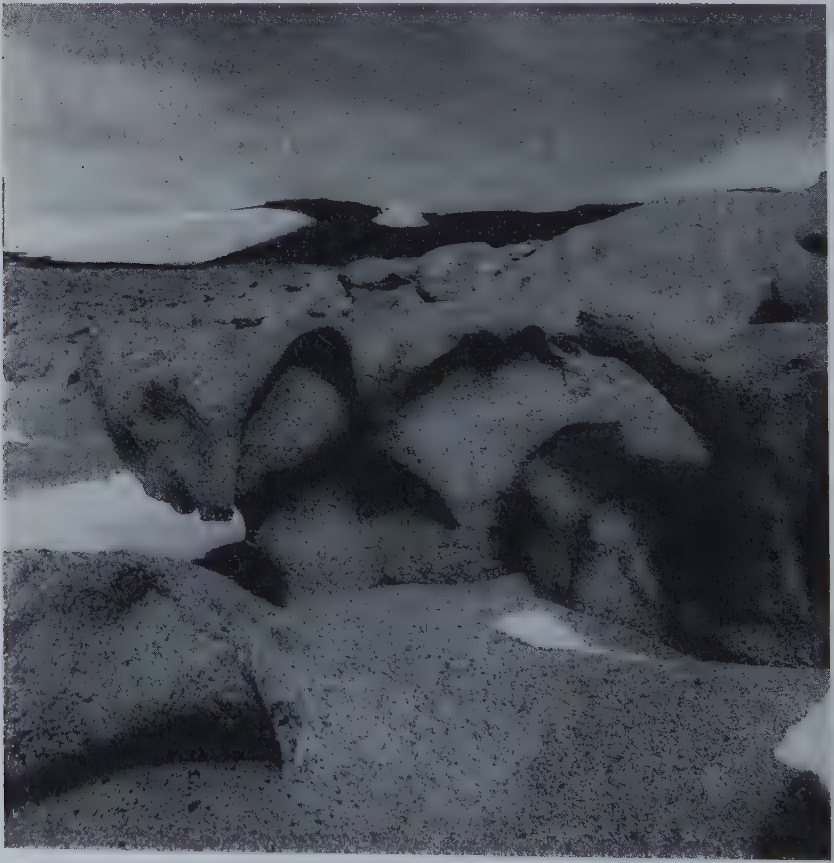

FIG. 12.—Granite, sculptured mainly by aeolian abrasion within a cirque of the Southern Range. Scale may be obtained from the dark inclusion (right centre) which is 9 inches in length.

DYKE INTRUSION

Petrography

The dykes consist of two types, pegmatites and lamprophyres. In hand specimen the pegmatite is a yellow rock commonly mottled by dark minerals which impart a foliation to the rock. Under the microscope it is seen to consist of a coarsely crystalline assemblage of orthoclase and microcline microperthite with a smaller amount of interstitial quartz. The yellow colouring is due to iron staining. In hand specimen the lamprophyre is a dark finely crystalline rock revealing in thin section phenocrysts of hornblende in a ground mass of plagioclase.

Structure

Pegmatite dykes are best exposed in the steep walls along the northern bordering range. The dykes maintain an average thickness of 9 to 12 ft and repeatedly bifurcate to fine stringers. At the surface they may weather out as elongate ridges up to 12 ft high which may be traced for several miles.

Lamprophyre dykes are common in the valley but are most concentrated on the 2,500 ft glacial shoulder below Mt. Insel (Figs 11 and 13). More than fifty NE striking dykes crop out over a distance of one mile, and about six NW striking dykes intersect with them. Each lamprophyre dyke maintains a uniform width of up to 12 ft along its entire length. Small stringers may bifurcate from each dyke but usually terminate within the granite. Small lamprophyre dykes, some only 3 inches in width extend close to and parallel with the thicker dykes.

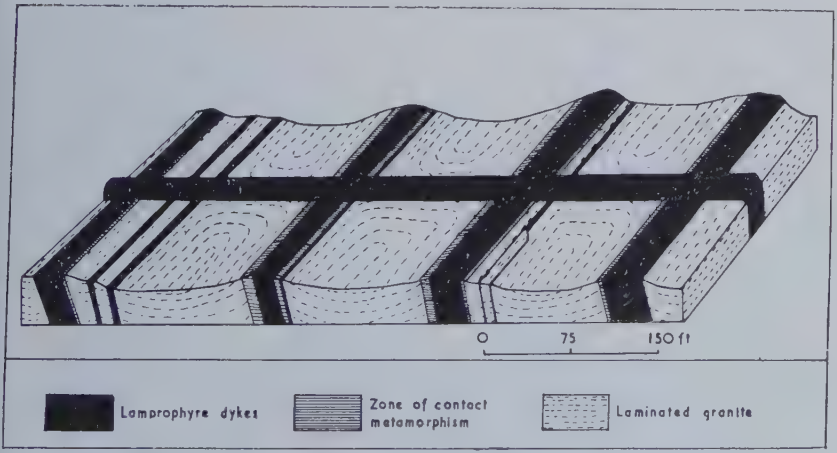

FIG. 13.—Diagram of a lamprophyre dyke swarm intruding granite. Along one or both margins of each dyke contact metamorphic schist or gneiss may be exposed. Adjacent parallel lamprophyre dykes are separated by troughs eroded in the friable laminated granite.

Contact Phenomena

The degree of contact metamorphism is approximately proportional to the width of the dykes. Along small dykes metamorphism is confined to a very narrow zone. In large dykes a marginal belt of granite almost as wide as the dyke itself is changed to low grade schist and gneiss. Common metamorphic rocks within the granite-lamprophyre association are quartz albite biotite, quartz plagioclase amphibole, and quartz plagioclase pyroxene schists and gneisses.

DOLERITE SILL INTRUSION

General Distribution

The dolerites of South Victoria Land extend over a wide area. They occur as dykes and sills in the basement complex, and form massive sills up to 2,000 ft thick in the overlying Beacon Sandstone. They also occur as boulders in moraines on the mainland and at Cape Royds on Ross Island.

Previous Work

Benson (1916) divided the dolerites into phaneric and aphanitic groups. The former have a medium grain size with an aphitic or gabbroid texture. The aphanitic rocks are very finely grained and have a granulitic structure. Among the different types Benson recognized were quartz dolerite, composed of basic plagioclase, monoclinic and rhombic pyroxene with varying amounts of quartz and feldspar occurring interstitially. The pyroxene occurred separately as hypersthene, enstatite and augite, or as enstatite and augite micrographic intergrowths.

The Dolerite of Victoria Dry Valley

Two extensive dolerite sills are exposed in the valley walls (Fig. 6); the upper now forming a remnant cap above the granite, the lower exposed as a continuous intrusion of uniform thickness through granite. The lower sill is approximately 1,400 ft thick and is separated from the base of the upper sill by 1,500 ft of homogeneous quartz, orthoclase granite.

The contact with the granite at the top of the lower dolerite sill was traced for several miles and found to be sharp and extremely regular (Fig. 6). This contrasts with the contact at the base of the same sill which is charterized by large finger-like extensions into the granite beneath.

Petrography

Microscopic examination has been made of samples collected at vertical intervals of 200 ft throughout the lower 1,400 ft sill. The principal minerals present are feldspar, mainly labradorite, and pyroxenes including augite, hypersthene, and pigeonite. Accessory minerals includes quartz, amphibole, apatite, biotite, iron ore, and chlorite Detailed study has confirmed the recognition in the field of two macroscopically distinct dolerite rock types.

One type occurs as two relatively thin near horizontal sheet-like bodies near the top of the sill. Its chararcteristic coarse grain facilitates its recognition in hand specimen. The texture may be described as hypidiomorphic granular, though there is occasionally a tendency towards sub-ophitic relations of pyroxene to feldspar.

Labradorite, An 60, comprises up to 60% of the rock. Individual crystals may exceed 4 mm in length and are usually well zoned. Some of the crystals are transected by fractures, the interstices of which are filled with aggregations of small grains of quartz.

Augite ($2V = 50°$, $Z \wedge C = 40°$) is the common pyroxene. The augite may be surrounded by a reaction rim of amphibole containing numerous inclusions of magnetite. A prominent schiller structure is displayed by the augite (developed parallel to (001)). Occasional crystals of pigeonite, some practically uniaxial, others with a small optic axial angle, have been observed. They are commonly surrounded by augite. Quartz occurs as graphic intergrowths with feldspar. The intergrowths are quite coarse and may comprise as much as 15% of the rock.

The other petrographic type, which forms the bulk of the sill, is generally finer in grain and consists almost entirely of plagioclase and pyroxene. Some samples are particularly feldspathic with labradorite forming 70% of the rock.

Quartz-feldspar intergrowths are rare in this type although quartz is present as a late forming mineral generally between the feldspar.

Hypersthene is the characteristic pyroxene ($2V = 80°$, En_{75}) occurring as large crystals with well developed exsolution structures. Augite is also present as are occasional grains of pigeonite. In one slide all three pyroxenes are present. Some sections consist of 55% pyroxene.

The relative proportions of ortho to clino pyroxene varies throughout this petrographic type with hypersthene being markedly predominant near the centre of the sill.

Jointing and Weathering

The dolerite of the upper sill is apparently more resistant to weathering than that of the lower. The columnar, angular or subangular jointed blocks of the upper sill are invariably fresh although covered with a thin ferruginous crust. The vertical columns of this sill form the sides and back walls of cirques within the Southern Range. The much coarser dolerite exposed in the lower sill weathers rapidly to honeycomb structures and subangular blocks which eventually crumble to a coarser grit.

CONCLUSION

The petrology of the granite, dolerite sills and dykes of Victoria Dry Valley resembles that of the Kukri Hills region to the south. Field relations are similar and it is probable that Victoria Dry Valley is part of the northern continuation of the basement complex of South Victoria Land.

ACKNOWLEDGEMENTS

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AXINITE FROM THE PERTH RIVER, WESTLAND, NEW ZEALAND

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(Received for publication, 25 November, 1958)

Summary

Axinite, associated with clinozoisite, occurs in quartz veins cutting sheared greywackes of the Chlorite 2 subzone in the upper valley of the Perth River. This is the first record of axinite in New Zealand.

INTRODUCTION

During Easter 1958, Dr Maxwell Gage and the author made a rapid reconnaissance of the Whataroa and Perth valleys, from the Alpine Fault eastwards almost to Scone Creek. The geological section is similar to that in the region farther south (Lillie and Mason, 1955): biotite gneisses at the Alpine Fault pass into less metamorphosed rocks to the east, until at the Main Divide the rocks are greywackes and argillites of the Chlorite 1 subzone. The material which forms the subject of this paper was collected in an unnamed stream which joins the Perth River on the south bank about a mile below the Scone Creek confluence.

DESCRIPTION

The country rock at this place is sheared greywacke of the Chlorite 2 subzone, and it contains numerous lenses and veins of quartz. These veins are quite narrow, seldom more than 2 in. thick. Most of them consist entirely of milky-white quartz. A few contain patches of blue-green pumpellyite. One of them showed a small amount of pale pink mineral that could not be identified in the field. Laboratory examination has shown that this mineral is axinite, which has not been previously recorded in New Zealand.

The axinite occurs as irregular grains and patches up to 5 mm across. It is associated in the quartz vein with pale grey clinozoisite. A thin

N.Z. J. Geol. Geophys. 2 : 137-40.

section of the vein shows anhedral axinite, slightly greyish and turbid from numerous tiny inclusions, clinozoisite in irregular prismatic grains, and quartz; the texture is xenomorphic-granular, and the three minerals appear to have crystallized more or less simultaneously. A small amount of interstitial calcite is present.

The optical properties of the axinite are as follows: $\alpha = 1 \cdot 674$, $\beta = 1 \cdot 681$, $\gamma = 1 \cdot 684$, optically negative, $2V = 70°$. The density, measured by suspension of small grains in an acetone-methylene iodide mixture, is $3 \cdot 18$; this figure is somewhat lower than that usually recorded for axinite, and the discrepancy may be due in part to the presence of quartz inclusions in the grains.

The formula of axinite can be written

$$(Ca,Mn,Fe)_3Al_2(BO_3)Si_4O_{12}(OH)$$

Analyses of axinite show that Ca:Mn + Fe is commonly about 2:1, and that up to about one-third of the (Mn + Fe) can be replaced by Mg. On account of these complexities, it is not possible to deduce the composition from the optical properties. The refractive indices of the Perth River axinite are as low as any recorded for this mineral, suggesting a considerable amount of replacement of (Mn + Fe) by Mg. The low density supports this suggestion.

The optical properties of the clinozoisites are as follows: $\alpha = 1 \cdot 713$, $\beta = 1 \cdot 720$, $\gamma = 1 \cdot 727$, optically positive, $2V = 88°$. It is non-pleochroic and shows normal interference colours. The optical data for the clinozoisite-epidote series, as given by Winchell (1951), are inadequate to derive an exact composition, but they indicate that the Perth River clinozoisite must be comparatively low in iron, with not more than about 10% of the $HCa_2Fe_3Si_3O_{13}$ component. The measured density is $3 \cdot 25$, which also agrees with a comparatively low iron content.

DISCUSSION

An examination of the recorded occurrences of axinite shows that it is usually found in one of the following associations:

(1) Contact-metamorphic rocks, often close to granite intrusions;

(2) Cavities in granites;

(3) Hydrothermal veins.

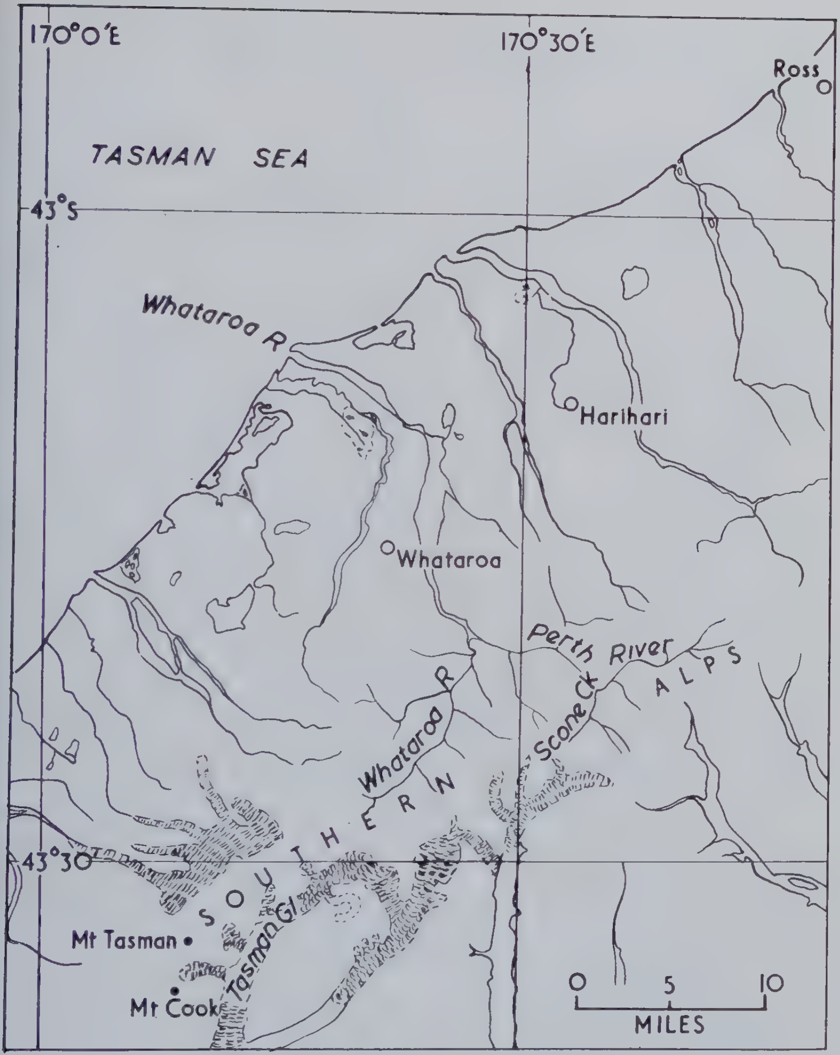

FIG. 1.—Locality map.

The presence of boron in axinite suggests the introduction of this element from a magmatic source, and the paragenesis and associated minerals usually indicate a comparatively high temperature of formation. These indications are, however, lacking in the Perth River occurrence. No granitic intrusions occur in the Southern Alps. The veins that occur in the greywackes and the schists appear to be segregations of quartz dissolved out of the country rock and redeposited in joints. The axinite-bearing vein penetrates sheared greywacke of the Chlorite 2 subzone. The vein has a sharp contact against the greywacke and does not appear to have affected the country rock in any way. This suggests that the solutions which deposited the vein minerals were little, if any, hotter than the country rock at the time of deposition. Under these circumstances, it seems that the components of the axinite, including the boron, were leached from the country rock by circulating solutions. The boron content of New Zealand argillites and greywackes is usually of the order of 100 to 800 p.p.m., much higher than the average crustal abundance of boron (3 p.p.m.). Tourmaline is not uncommon as an accessory mineral in the schists and gneisses of the Southern Alps, and this tourmaline is to be ascribed to the boron content of the original rock (Reed, 1958). It seems reasonable that the material forming the axinite was also derived from the enclosing greywackes and argillites.

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THE HOT SPRINGS AND HYDROTHERMAL ERUPTIONS OF WAIOTAPU

By E. F. Lloyd, New Zealand Geological Survey, Department of Scientific and Industrial Research, Rotorua

(Received for publication, 6 November 1958)

Summary

On the basis of chemical composition, the waters of hot springs of the Waiotapu thermal area have been divided into four classes: acid sulphate waters, sulphate-chloride waters, chloride waters, and bicarbonate-chloride waters. The origin of the waters and the relationship of the hot springs to geological features is discussed. The natural heat flow of the Waiotapu thermal area has been estimated as 302,000 Kcal/sec above 15° C.

About 900 years ago hydrothermal eruptions took place throughout the Waiotapu district, and these are discussed especially in relation to the close association of present thermal activity with craters formed during these outbursts, and the effects of the volcanicity on the hydrothermal system. A possible mechanism for triggering these eruptions is advanced.

INTRODUCTION

Waiotapu is 15 miles south-south-east of Rotorua and 25 miles north-east of Wairakei, on the Rotorua-Taupo State Highway (State Highway 21) (Fig. 1). The thermal area is contained in sheets N85/3 and N85/6 of the N.Z.M.S. 86 Series (Fig. 2). The area was by-passed by early travellers who followed the track between Rotorua and Taupo

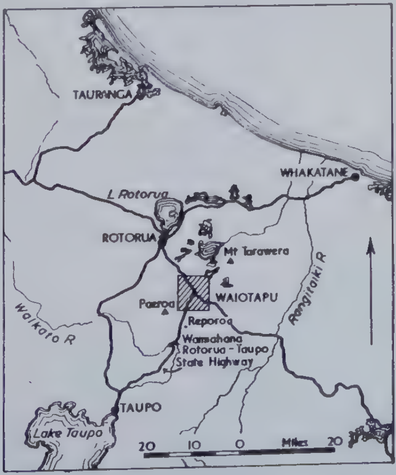

Fig. 1.—Locality map of Waiotapu district.

along the western side of the Paeroa Range. It was only after the Pink and White Terraces were obliterated by the Tarawera Eruption in 1886 that Waiotapu became prominent as a tourist resort. The first descriptions of Waiotapu appear in reports on the eruption of Tarawera and Rotomahana (Smith, 1887; Leys, 1887; Thomas, 1888).

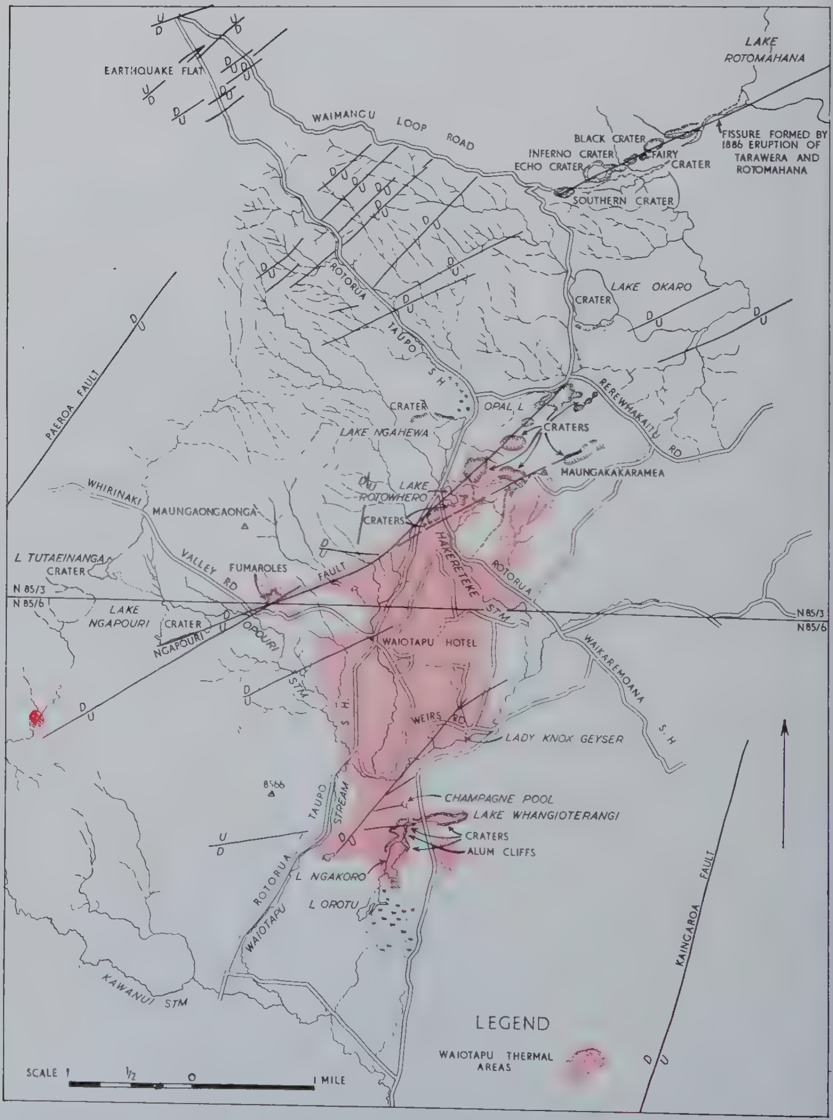

FIG. 2.—Map showing main faults, division between sheets N85/3 and N85/6, and extent of Waiotapu thermal area.

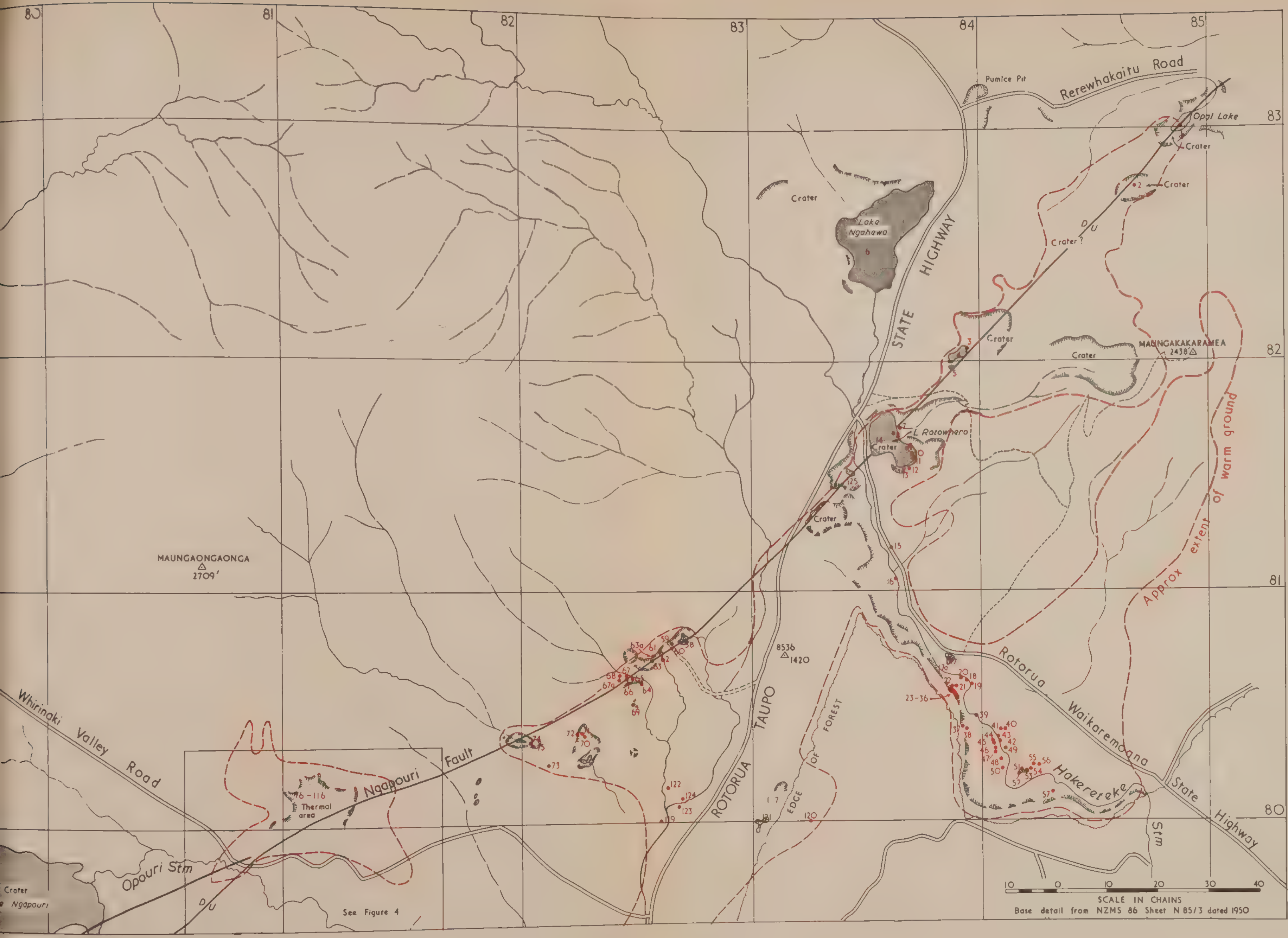

FIG. 3.—The Waiotapu hot springs of Sheet N85/3.

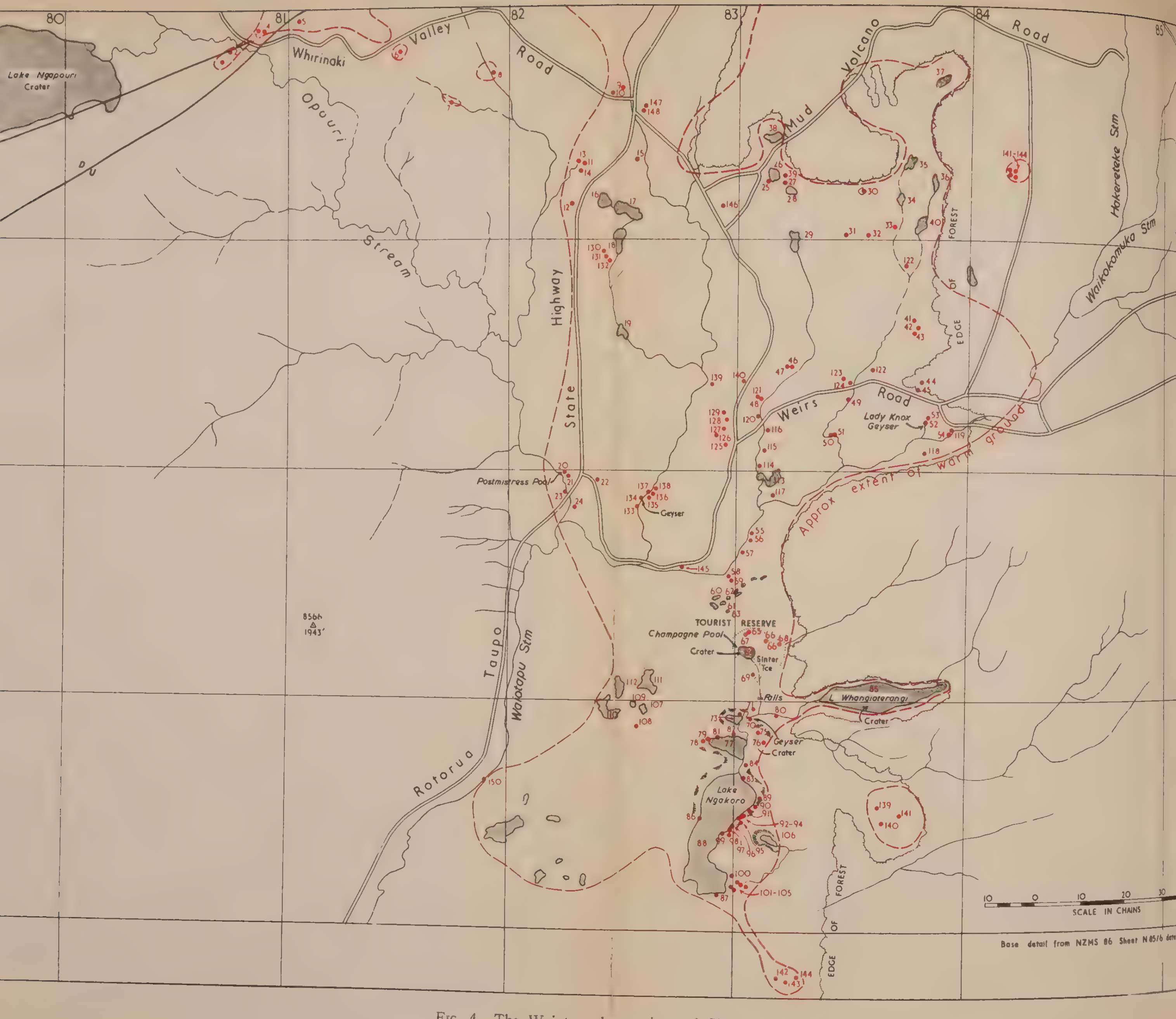

Fig 4.—The Waiotapu hot springs of Sheet N85/6

Herbert (1921) discusses the balneological properties of some Waiotapu spring waters. The most important account is that of Grange (1937), who mapped and described the main springs.

The present field survey was commenced late in 1955, and was completed in November 1957. The hot springs (Figs 3, 4, 5) have been mapped, and a system introduced by Messrs D. R. Gregg and A. C. M. Laing* (pers. comm.) of numbering the springs separately in each sheet of the N.Z.M.S. 86 Series has been followed, so that those in the northern part of the area are in sheet N85/3 (Fig. 3), and those in the southern part in sheet N85/6 (Fig. 4). To avoid confusion spring numbers are prefixed by sheet numbers.

Hot spring waters were analysed by the writer for chloride, fluoride, bicarbonate, carbonate, and boric acid; pH and electrical conductivity were also determined. The natural heat flow from the Waiotapu thermal area was estimated by Mr R. F. Benseman and the writer. The work was carried out primarily in connection with Waiotapu geothermal investigations.

TOPOGRAPHY AND GEOLOGY

The Waiotapu thermal area is at the northern end of the Reporoa lowland which occupies a fault angle depression between the Kaingaroa Fault (Grange, 1937, p. 49-50) on the east and Paeroa fault-block on the west. The depression extends from Lake Taupo in the south to the twin volcanoes Maungakakaramea (2,438 ft), more commonly known as Rainbow Mountain, and Maungaongaonga (2,709 ft) in the north. There are six craters on the steaming slopes of Maungakakaramea, and small lakes occupying the sites of explosion craters are a conspicuous feature of the Waiotapu landscape. These craters are described in detail later. One mile south of Maungaongaonga is a large extrusive dome (Trig 8566, (1,934 ft)).

The area is drained by the Waiotapu Stream which flows south, joining the Waikato River at Waimahana, near the Mihi Bridge. Hakereteke Stream (Kerosene Creek) has its headwaters north-west of Maungakakaramea and flows through the thermal area to join the Waiotapu Stream. Opouri Stream drains Lakes Ngapouri and Tutaeinanga in the depression on the crest of the Paeroa block, and flows approximately south-east to join the Waiotapu Stream south of Waiotapu Hotel.

Kaingaroa Fault is 1 mile east of the thermal area, and has been masked by a younger ignimbrite flow which extends west to the edge of the thermal area and north to near Maungakakaramea. South-east of

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*Hot Springs of Sheet N94/4, with Notes on Other Springs of N95. In "Results of Geothermal Investigations, Wairakei Taupo Areas". Unpublished D.S.I.R. Report on open file, D.S.I.R., Wellington.

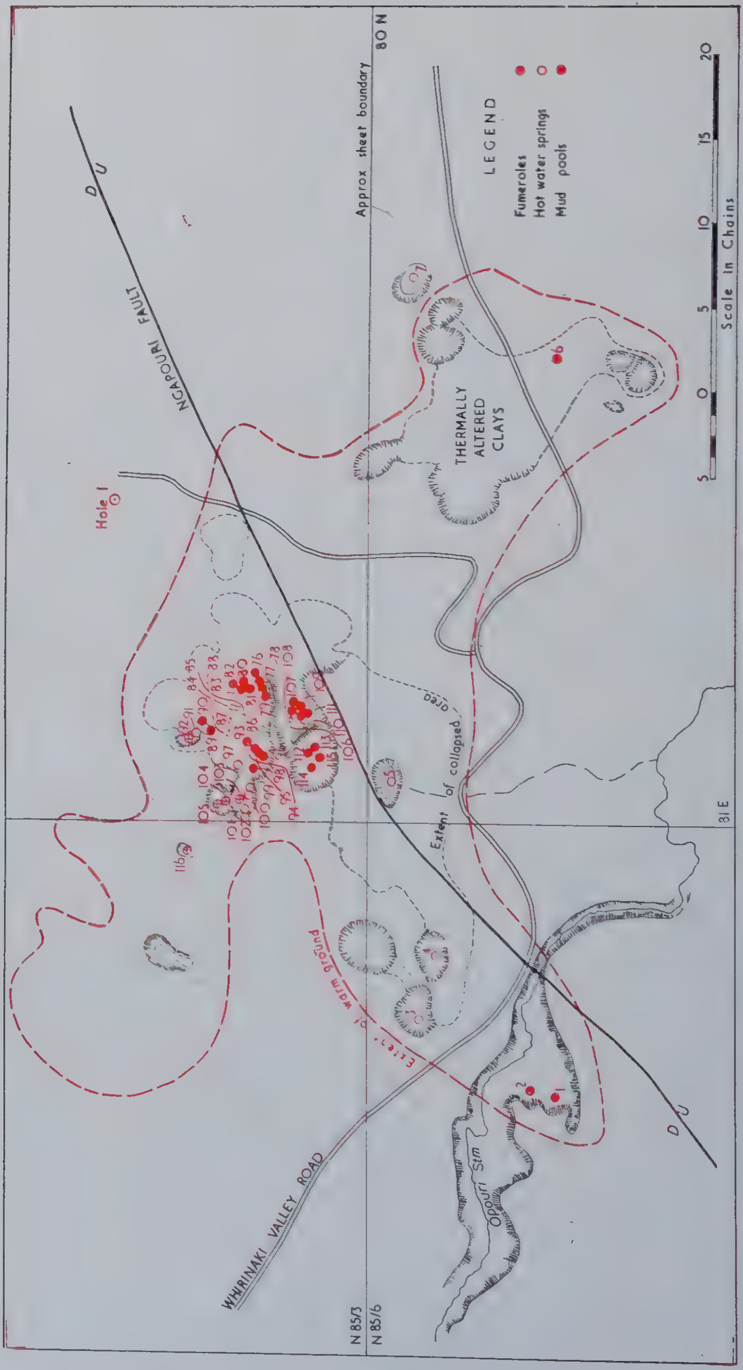


FIG. 5.—The hot springs of the Maungaongaonga fumarolic area.

the thermal area the Kaingaroa Fault forms a prominent scarp along the west edge of the Kaingaroa ignimbrite plateau, and its continuation southwards forms the eastern margin of the Reporoa lowland. On the west rise the backslopes of the Paeroa block to an altitude of 3,212 ft, terminated abruptly by the steep scarp of the Paeroa Fault. This trends north-east and has maximum downthrow of 1,000 ft to the west. East of it the block is crossed in an east-north-east direction by the subsidiary Ngapouri Fault, which branches from the former fault near Paeroa Trig (3,212 ft) and cuts across the lower south-eastern slopes of Maungaongaonga and the north-west flanks of Maungakakaramea. Ngapouri Fault is downthrown approximately 200 ft to the north-west and along it are recent craters. In some of these thermal activity still exists. This fault is named after the largest of the craters, now occupied by Lake Ngapouri.

Steiner (1958) has re-classified as dacite the rocks of Maungaongaonga, listed as acid andesite by Grange (1937). The rocks of Maungakakaramea, also classified as acid andesite by Grange, closely resemble Maungaongaonga rocks in hand-specimen, and are probably also dacite. The rocks of the dome, (Trig 8566) are rhyodacite.

The surface of the area is mantled with poorly compacted volcanic ash, the youngest of which is pumice erupted from Taupo about 130 A.D. (Baumgart, 1954; Baumgart and Healy, 1956; Fergusson and Rafter, 1957). This pumice has been covered by hydrothermally altered country rock erupted from craters along Ngapouri Fault, at Lakes Ngakoro and Whangioterangi, at Alum Cliffs, Champagne Pool, and north-west of Champagne Pool. On the flood plain of Waiotapu Stream, below its junction with Opouri Stream, the Taupo pumice is covered by alluvium derived from the young hydrothermally altered breccias, and was probably eroded from the slopes of Maungakakaramea, since it includes fragments of dacite. Rotomahana Mud, erupted during the Tarawera eruption of 1886, forms a thin cover over the young hydrothermally altered material on the lower north-eastern slopes of Maungakakaramea, but does not extend farther south.

At Alum Cliffs, horizontally bedded lacustrine sediments about 100 ft thick, are exposed beneath the volcanic ash beds. These sediments are composed of rhyolitic pumice and other water-sorted pyroclastic material (mapped by Grange, 1937, as Waitahanui Breccia) and are probably outwash from the pumice breccias whose source is near Earthquake Flat. They extend south to Waiotapu as a large fan. Locally these beds have been silicified by hydrothermal action. South of Alum Cliffs they are elevated above the general level of the surrounding land and could be a remnant of an old land surface, but appear to be arched up, perhaps by intrusion. The Ngakoro-Champagne Pool area coincides with a minor high gravity anomaly (Mr F. E. Studt, Geophysics Division, D.S.I.R., pers. comm.), and this strengthens the hypothesis for intrusion. This elevated block has explosion craters about its base; the largest and best preserved is now occupied by Lake Ngakoro. The formation of this crater has resulted in the greater part of the eastern portion of the block being blasted away.

Several sets of lake terraces have been cut around the Reporoa lowland, between the 1,000 ft and 1,500 ft contours. The upper terraces are overlain by the ash showers and are probably of late Pleistocene age. A strand line of pumice blocks at approximately 1,070 ft indicates the maximum level reached by a lake which flooded the lowland following the Taupo pumice eruptions of 130 A.D. This level was maintained for a very short period as there is no evidence of wave cutting, but a lake at 1,000 ft probably existed for a longer period (Mr C. G. Vucetich, Soil Bureau, D.S.I.R., pers. comm.).

Hot springs, fumaroles, and warm ground occur within an area of six square miles, on an eroded fan extending south from Maungakakaramea for a distance of 3½ miles, and on the south-eastern slopes of Maungaongaonga, and on the west and south flanks of Maungakakaramea. The area, approximately triangular in shape, is bounded by the Kaingaroa State Forest on the east, Ngapouri Fault on the west, and Lake Ngakoro on the south, with Maungakakaramea forming the apex.

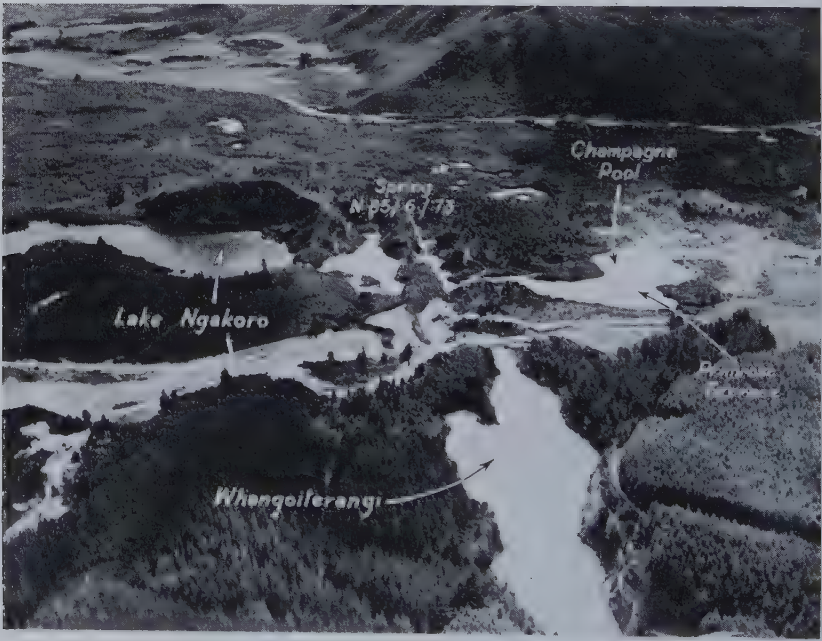


FIG. 6.—Air view looking west, showing crater occupied by Lake Ngakoro (centre left), Whangioterangi (foreground) and the small craters at Alum Cliffs (centre). The "Roof Mountain" through which Ngakoro Crater has been blasted is clearly shown, and also the alignment of the Whangioterangi and Alum Cliffs Craters. Spring N85/6/73, the source of the large sulphur globules, occupies the western-most crater in line with the long axis of Whangioterangi. Champagne Pool, spring N85/6/64, and the Primrose Terrace are at the right. In the background rises the rhyodacite dome, Trig 8566.

Photo: National Publicity Studios.

Within this area, lines of elongated explosion craters at Alum Cliffs (Fig. 6) are considered to be located on faults. They strike at approximately 080°, and intersect another zone on which many hot springs occur. This zone trends north-north-east from just west of Lake Ngakoro to the north of Weir's Road and is considered to lie along another fault. The fault traces have been obliterated by the alluvium and eruption products that fill this part of the Reporoa lowland.

DESCRIPTION OF THE HOT SPRINGS

The springs previously described by Grange (1937, pp. 94-6) fall into two groups. One group comprises those situated along five miles of the Ngapouri Fault (Fig. 3), including from S.W. to N.E. several small areas where gas escapes in the depression on the crest of the Paeroa block, the area of intense fumarolic activity on the south-eastern slopes of Maungaongaonga, the springs in the small explosion craters farther north-east to Lake Rotowhero, and the steaming ground on the north-west flank of Maungakakaramea. Springs at the southern base of Maungakakaramea on the banks of Hakereteke Stream are also included in this group because they are similar chemically to the springs at Rotowhero.

The other group (Fig. 4) extends south for 3½ miles from Maungakakaramea to Lake Ngakoro and includes the main Waitapu thermal area visited by tourists. The springs become more numerous towards the south, and at the extreme south of the area are the explosion craters mentioned previously.

All the springs, except those along the Ngapouri Fault, are close to surface water, and flowing springs occur between the 1,050 ft and 1,600 ft contours. Steaming ground extends almost to the summit of Maungakakaramea.

Chemical analyses of the spring waters are given in Table 1.

In the Reporoa lowland, south of Waiotapu but outside the area of the present survey, are several small, isolated groups of hot springs. Some of these springs contain appreciable quantities of arsenic (Grimmett and McIntosh, 1939, pp. 137-45).

Springs on Ngapouri Fault

ACTIVITY ON THE CREST OF PAEROA BLOCK

South-west of Lake Ngapouri (approximate grid ref. 774779) are several small areas where present-day activity is confined to the escape of gas and deposition of sulphur (Fig. 1). There are jets of steam in one area, but the surface is cold in the remainder. Widespread sulphur deposits indicate that within quite recent times the activity has been more intense.

Adjacent to these areas Taupo pumice rests on extensive beds of siliceous sinter. The pumice fragments are not silicified or cemented to the sinter, so the thermal activity which formed the terraces was probably extinct when the pumice fell. Judging by the extent, thickness, and structure of the sinter, the springs that formed them had a copious discharge of hot alkaline water. Many plant impressions are preserved in the sinter.

It is suggested that the depression was drained in a southerly direction by the Kawanui Stream, and that swampy conditions or shallow lakes existed, indicating a plentiful supply of ground water and alkaline conditions at the thermal area. At some time prior to the Taupo Pumice eruptions (about 130 A.D.), Opouri Stream captured the drainage from the head of the depression, lowering the level of the water table in the vicinity of the thermal area, and causing a change from alkaline to acid activity, but little change in heat output. The subsequent eruptions from nearby craters (Lakes Ngapouri and Tutaeinanga) may have released sufficient energy to cause a marked decline in the thermal activity at this locality.

MAUNGAONGAONGA FUMAROLIC AREA

Fumaroles, steaming ground, and hot springs extend over an area of ten acres on the south-eastern flanks of Maungaongaonga, and on the southern flanks of the mountain warm ground extends well up some of the ridges (Fig. 5). With the exception of spring N85/3/102, all the waters are acid and have very low mineral content. The most vigorous fumaroles and the only flowing springs are in the intensely active gulch near the centre of the area. The fumaroles are all superheated, and temperatures between 99·8° C and 102° C (spring N85/3/105) have been recorded. The flowing springs (N85/3/99-102) are all near the head of the gulch, and their combined discharge forms a small rill which flows for 30 yards or so and disappears into a collapse pit.

Spring N85/3/102 seeps from the steep bank behind spring 101. The water has a pH of 7·4, and white sinter stalactites are being formed. Apparently this water is supplied from a perched water table, and becomes heated in passing through the thermal area.

Below the gulch are numerous mud volcanoes which depend on rainfall for their water supply.

The Maungaongaonga springs are perched high above the water table; measurements in Ministry of Works drillhole No. 1 indicate that the water table is about 300 ft below the surface. The spring waters contain very little mineral matter, and are of acid composition, presumably of the acid-sulphate type. Because the springs lie well above the main body of ground water, little information can be gained from them as to the type of water at depth, though it is considered likely to be of the chloride type.

All the springs and fumaroles are confined to the downthrown side of Ngapouri Fault, but steaming ground extends on both sides (Fig. 7). The Maungaongaonga fumarolic area has the highest heat output per unit area at Waiotapu.

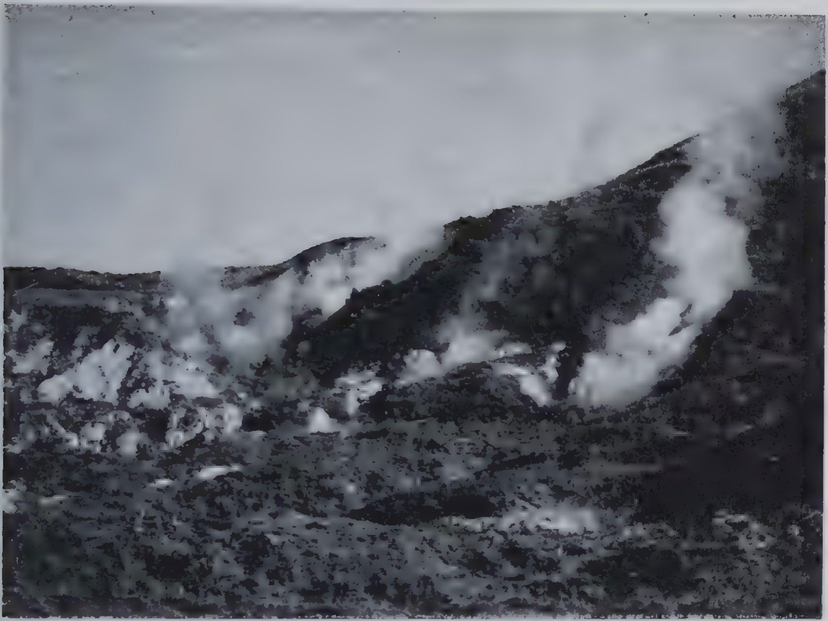


FIG. 7.—Maungaongaonga fumarolic area. The steam column on the right is from the fumarole, N85/3/92, and the fumaroles in the active gulch can be seen near the centre. The mud volcanoes are at the bottom left, and Ngapouri Fault runs across the foreground of picture.

SPRINGS BETWEEN MAUNGAONGAONGA FUMAROLIC AREA AND MAUNGAKAKARAMEA

For half a mile to the north-east of Maungaongaonga fumarolic area, Ngapouri Fault has a discontinuous trace with recent downthrow approximately 3 ft to the north-west, but there is no thermal activity. Farther north the fault is marked by a series of explosion craters, some of which contain hot springs (N85/3/58-63). A notable feature of the springs in this area is that most have a temperature of about 50° C, and a pH of about 3·2, indicating that they are supplied by a uniform body of ground water. Hole 2, recently drilled in this area, has a permanent water table at approximately 70 ft, so it is likely that the ground water supplying these springs is perched. Exceptions are springs N85/3/59, 63, and 66. Springs N85/3/59 and 63 have temperatures of 65° C, and spring N85/3/66, which is in the bed of the stream formed by the discharge from springs N85/3/64-68 where it is crossed by Ngapouri Fault, has a temperature of 99° C several inches below the stream bed. These springs lie directly on Ngapouri Fault and probably owe their high temperature to steam being jetted up the fault plane. Springs N85/3/70-72 have small deposits of limonite in their overflow channels. All the springs depend on rainfall for their water supply, and in some summer months cease to flow. The waters are similar in chemical composition to those of the Maungaongaonga fumarolic area.

MAUNGAKAKARAMEA

These springs are shown in Fig. 3. Lake Rotowhero, at the junction of the Rotorua-Taupo and Rotorua-Waikaremoana State Highways, occupies one of the craters on Ngapouri Fault. Rotowhero is coloured bright green by algae and has a temperature of approximately 30° C. It is fed by hot springs, N85/3/8-13, on its north-eastern shore. These are at or near boiling point, and each discharges about 1·5 litres per second of highly acid water (pH 3·0) containing approximately 200 parts per million of chloride ion. The waters of this area are chemically similar and probably originate from a common source.

North from Rotowhero, and separated from it by only a narrow ridge, is another large crater containing two small lakelets—one hot (spring N85/3/4), and the other cold (spring N85/3/5). Both occupy depressions within the crater and probably mark the sites of former vents. During the winter months a small rill of hot water (spring N85/3/3) flows from the stony fan at the northern end of spring 4. These waters have a pH of 2·0. The floor and sides of the crater are steaming and in places sulphur is being deposited. On the northern wall there as a horizontal grey band, thought to be a high water mark, which may indicate that water once filled the crater.

There are no other hot springs on the slopes of Maungkakaramea but almost everywhere the ground has been altered by steam action, which is now most vigorous on the southern slopes of the mountain, extending in patches from the base almost to the summit. The bare northern slopes adjacent to the craters on Ngapouri Fault are still warm, but have been more active in the past. Hochstetter (1864, p. 144), who viewed the northern slopes of Maungakakaramea in 1859 from Lake Rotomahana, describes the mountain thus: "On its fern and shrub covered slopes many open places can be seen, on which red coloured soil is exposed to view, and in all these places from foot to summit of the mountain, steam escapes so that the mountain is often quite enveloped in steam as if it were a volcano."

SPRINGS AT HAKERETEKE STREAM

Hot springs, N85/3/15-57 occur on both banks of Hakereteke Stream where it skirts the southern base of Maungakakaramea. Chemically the springs are similar to those at Rotowhero, but petroleum occurs in some small springs, N85/3/23-36, at the foot of a steep, bare face on the right bank of Hakereteke Stream (grid ref. 839805), where there is a strong smell of hydrogen sulphide and crude oil. The petroleum collects on the surface of the non-flowing springs, and a chocolate-brown coloured deposit of ozokerite surrounds the springs. Morgan (1922, p. 13) suggested that the petroleum may be formed by distillation of carbonaceous material included in the ignimbrites and volcanic tuffs, and Park's hypothesis (1923, pp. 135-6) was that it was being distilled from coal-bearing beds similar in age to the low grade coals found in the Waikato district. The author considers Park's hypothesis unlikely, and Morgan's probably correct, though not as to the source of the carbonaceous material. Lacustrine sediments similar to the mudstone of the Huka Formation at Wairakei (Steiner, 1953) underly the area, and the author suggests that the oil is being distilled from carbonaceous material in these sediments. It may be possible to verify this if drilling operations are undertaken here.

On the edge of the forest, three-quarters of a mile north-east of the Waiotapu Hotel (grid ref. 802832), there is a small crater-like hollow containing numerous steam jets and shallow, bubbling, or "frying-pan" type of hot springs. There is no water discharge from the crater, and during long spells of dry weather the springs dry up, giving way to steam jets. Because the activity frequently changes its position within the crater the springs have not been numbered individually but are grouped as spring N85/3/117.

Main Waiotapu Springs (Fig. 4)

On the high ground in the forest at the north of the area the springs are acid, evolve much gas, and have small overflows. Immediately south of the plantation boundary there are several muddy acid springs, N85/6/26, 27, 28, 30, 34, 35-40. Spring N85/6/25 (*Te Huinga*) is the largest mud pool in the Waiotapu district, and until about 1925 had built a mud cone up to 20 ft high (Fig. 8). Since then the cone has been eroded away, presumably because the spring decreased in activity and ceased depositing fresh mud on the outer slopes. Springs

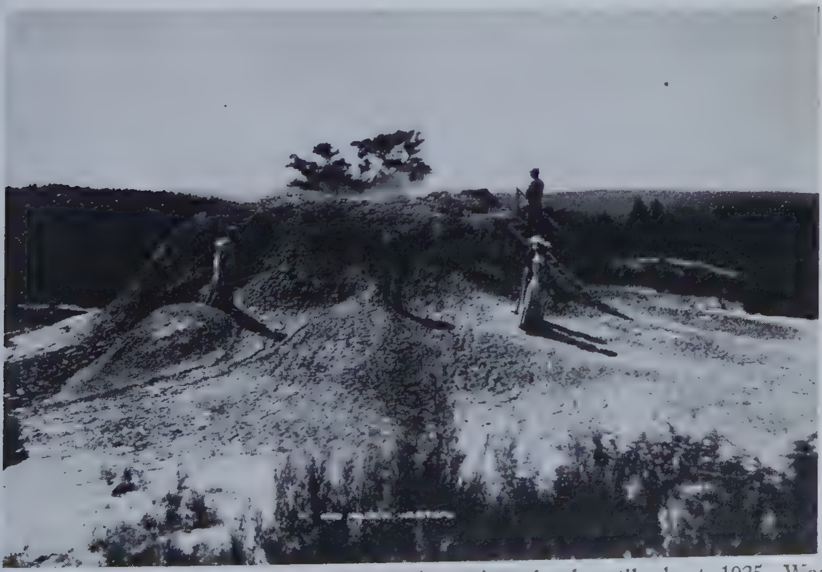


FIG. 8.—Mud Volcano, spring N85/6/25, as it existed until about 1925. Was the largest mud cone known in New Zealand thermal areas, but has since been eroded away. *Photo* (1910): *C. Spencer.*

N85/6/27, 30, and 39 are mud pots, evolving much gas. Grange mentions a spring in this locality which erupted muddy water to a height of 30 ft or more, and it is reported in the "New Zealand Herald" (August 7, 1905) that a party of tourists was covered in mud by an eruption which reached an estimated height of 160 ft. There are no records of similar activity recently, and the location of the erupting spring cannot be accurately pin-pointed from the old descriptions, but it is thought to be either spring N85/6/27 or 39. Springs N85/6/26 and 28 are shallow ponds of muddy water; each contains many vents, and periodically discharges gas and mud violently through the cooler surface water. The mud pools generally have no overflows but become violently active at times, usually during periods of low rainfall. During these active periods mud is thrown high in the air, building cones around the vents.

Farther south, at a lower level, and aligned nearly east and west on approximately the 1,200 ft contour, there are chloride springs, N85/6/13, 15, 29, 31, 32, and 33, all having discharges between 1 and 5 litres per second. These springs are the sources of the streams which flow south through the thermal area and join Waiotapu Stream near the Tourist Reserve.

South of the above line, the largest springs, at or near stream levels, discharge alkaline waters with chloride contents of 200 to 1,800 p.p.m. There are sulphate springs on the higher ground between the streams.

Lady Knox (spring N85/6/52) is an artificial geyser constructed about 1906 by Mr Scanlon, then chief warder of the Waiotapu prison farm (Fig. 9). When first reported by the press (*N.Z. Herald*, 6 December 1906), the geyser was referred to as the "Northland Geyser", and in later tourist publications as the "Waiotapu Geyser", but the present name of Lady Knox Geyser has become firmly established. Little information can be traced concerning the original physical condition of the spring, but it is thought to have been a vigorously boiling chloride spring with moderate water discharge. Following the discovery that the spring would erupt when soaped, Mr Scanlon restricted the size of the orifice with a cast iron pipe, to increase the height of the geyser displays, and camouflaged the pipe by building around it a cone of pumice blocks which have since been covered with sinter, giving the structure an almost natural appearance. Lady Knox Geyser does not overflow between eruptions and will not erupt unless it is soaped. It takes but a few seconds to respond to soaping, ejecting a thin column of water to heights of up to 50 ft. Often it will continue in eruption for several hours.

In the past there have been many alkaline springs in the area adjacent to Lady Knox Geyser. One chain south-east of the geyser there are several old sinter cones, showing no evidence of recent activity. West of the geyser are chloride springs N85/6/49, 50, and 51, on the banks of the stream draining a hot lakelet, spring N85/6/39.

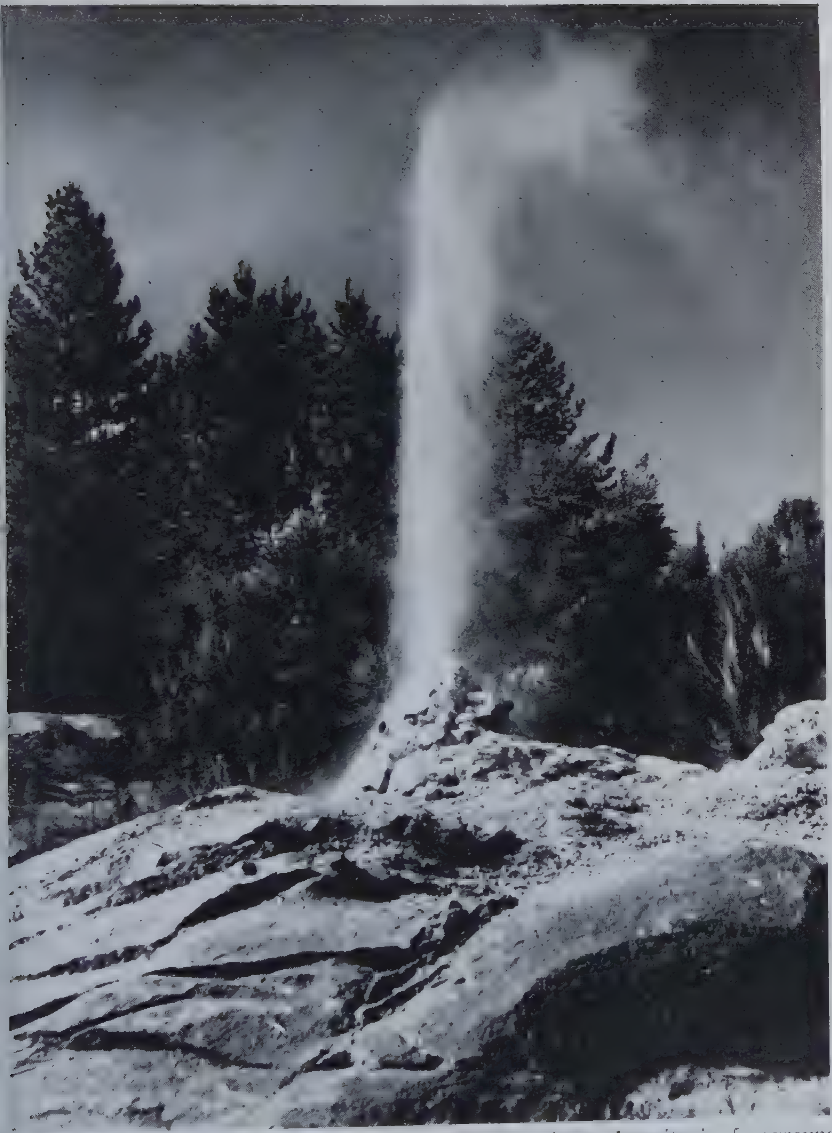


FIG. 9.—Lady Knox Geyser in eruption. Siliceous sinter deposits in foreground are typical of the sinter terraces formed by past chloride springs.

On the left bank of Waiotapu Stream there is a once extensive sinter terrace, now fissured, with sinter blocks tilted at all angles. On it are three sinter cones, the largest about 3 ft high, and several hot spring basins, now dry. Chloride water discharges on the banks of the Waiotapu Stream from springs N85/6/55, 56, and 57, beneath the sinter sheet. Possibly these are the springs which originally deposited the sinter and then broke out at a lower level as a result of the stream cutting down through the sinter sheet and lowering the surrounding water table.

Venus Bath (spring N85/6/48), a large chloride spring approximately 20 ft diameter on the bank of the stream draining spring N85/6/29, now discharges a mere trickle. Massive sinter deposits at the overflow suggest this spring was more active in the past. A few chains north of Venus Bath, the acid spring N85/6/46 is depositing a small terrace of sulphur, known as *Sulphur Terrace*.

Springs N85/6/133-138 lie in a gully to the west of the guide's house (and the site of the old Waiotapu Hotel), on both banks of the stream. They discharge an insignificant amount of chloride water, and deposit a greyish-white sinter. Mr R. F. Keam (pers. comm.) observed eruptions of springs N85/6/133 and 134 in April 1952. The former then had a period of 47 seconds with a 15 second eruption to 4 ft. When observed by the writer in December 1956, eruptions reached a height of 2 ft, but during the following winter the spring was discharging quietly. Spring N85/6/134 was not observed in eruption during the present survey.

The Postmistress (spring N85/6/20), close to Opouri Stream near the bridge on the Rotorua-Taupo State Highway, often shows a degree or so of superheat just below the surface, and discharges chloride water.

The main features of the Tourist Reserve are the crater-like pits and the Champagne Pool (N85/6/64). The pits occur along a north-east trending line which is interpreted as fault controlled. Due to intense thermal alteration of the ground about these pits, it is not possible to deduce whether they are due to collapse or explosion, but some at least are thought to have had an explosive origin. *Wharepunui* is a pit formed early this century by collapse, and has fine examples of sulphur crystals on its sides. *Katuarehi* (spring N85/6/63) is a violently boiling pool of muddy water at the bottom of one of the craters, and occasionally erupts to a height of 60 ft. This spring was reported in eruption in September 1902 ("New Zealand Herald", Sat. September 13th, 1902) and it was then stated that none of the Maori inhabitants remembered it being active previously. To the writer's knowledge, it was last in eruption during November 1953. Several other pits contain pools of hot water, springs N85/6/60-62, but the majority are now dry, though some are alleged to have been filled with water prior to the Tarawera Eruption in 1886. All the springs are acid, and being confined to the bottom of deep pits, have no surface discharge. A stream of near boiling water emerging from spring N85/6/59, at the head of the gully on the north of the track from the bridge across the Waiotapu Stream, near the guide's house, may represent the water discharge from the craters. The craters are up to 60 ft in diameter and 40 ft in depth and sulphur is deposited by the vapours escaping from the creviced sides and floors.

Champagne Pool (N85/6/64) on the flat area south-east of the pits, s one of the largest springs in the district, being fully 3 chains in diameter (Fig. 10). Its water has encrusted an area of about 3 acres with cream coloured sinter, known as the *Primrose Terrace*. The spring has a raised sinter rim on all sides except where it overflows on the east, and on the western side this rim raises water level about 1 ft higher than the surrounding ground. The overflow normally flows away to the south-east over Primrose Terrace, but deposition of sinter to the east of the spring is causing an ever increasing volume to flow away to the west. This is particularly so during easterly or south-easterly winds, when Primrose Terrace is often dry. The spring has a temperature of approximately 74° C, evolves considerable quantities of gas, and contains 1,800 p.p.m. of chloride ion. Primrose Terrace terminates at the *Bridal Veil Falls*, below which the discharge from Champagne Pool joins the discharge from *Whangioterangi* (Echo Lake, spring N85/6/85), and after meandering across several partly infilled explosion craters at *Alum Cliffs*, flows into *Lake Ngakoro* (Fig. 6).

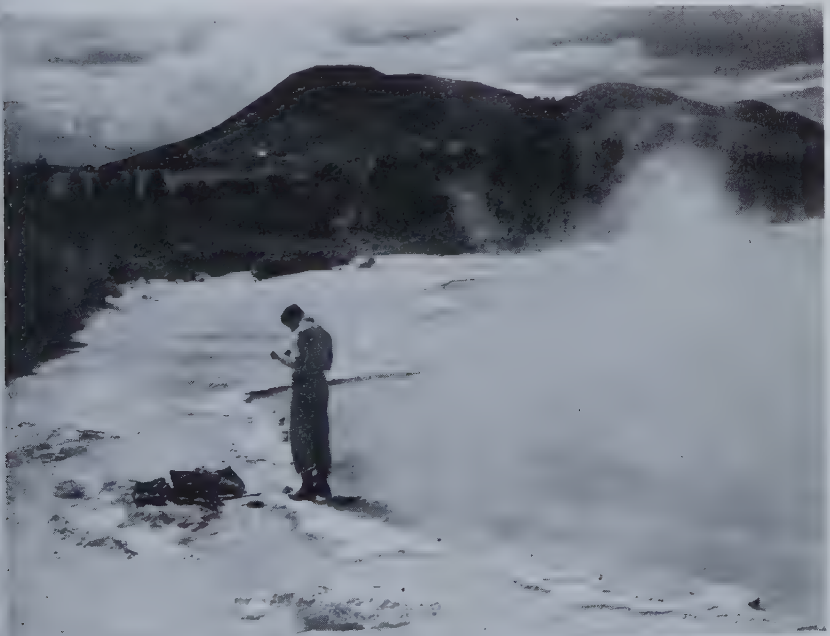


FIG. 10.—Champagne Pool, N85/6/64. Note the raised sinter rim and gas bubbles persisting on water surface. Dacite volcano Maungaongaonga in background, with steam showing from the fumarolic area on its south-eastern flanks.

WHANGIOTERANGI (ECHO LAKE) AND ALUM CLIFFS

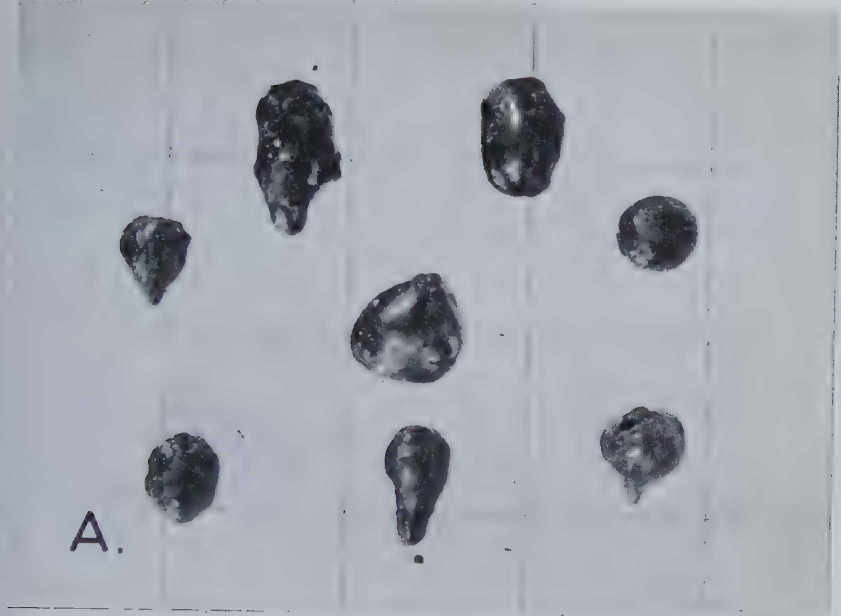
Whangioterangi is an oval shaped crater lake, 26 chains long and 6 chains wide, approximately 20 chains south-east of Champagne Pool. It is enclosed by high vertical cliffs on all but the western side where it is drained by a small stream which flows west-south-west for 20 chains and then south to join Lake Ngakoro. The course of the stream is determined by a number of explosion craters which, together with Whangioterangi, lie on a line trending at 080° which the writer considers to be a fault (Fig. 6). The craters south-west of Whangioterangi have been partly infilled, but their precipitous walls, the Alum Cliffs, remain to mark their outline. There are hot springs on the crater floors, and aluminium sulphate and other salts are deposited on the cliffs by water percolating through the exposed lacustrine sediments.

The occurrence of natural black sulphur is rare in New Zealand thermal areas, but has been reported by Hector (1887, p. 442), Thomas (1888, p. 10), and Grange (1937, p. 127-8) from Whangioterangi. Keam (1955, p. 29) states that Whangioterangi was disturbed by subterranean activity during 1948. In October 1954, following a brief period of seismic activity, gas was observed by Mr B. N. Thompson rising from two locations in Whangioterangi. "Near the northern edge and about mid-way along the lake, gas was rising at a moderately rapid rate over an area of about 10 yards in diameter, and about 5 chains east of the western end of the lake over an area of about 5 yards in diameter. Many black hollow spheres and pear-shaped globules about $\frac{1}{4}$ in. in diameter were floating on the surface of the lake, and were stranded round the shore line. At the place of maximum gas ebullition the depth was 80 ft, surface temperature 23° C, and bottom temperature 26° C. Where gas was rising at the eastern end of the lake, the depth was 52 ft and the bottom temperature was less than 24° C. The temperature of the discharge from the lake was approximately 23° C." (Mr B. N. Thompson, N.Z. Geological Survey, pers. comm.)

The outside surface of the globules was smooth, with a shiny metallic lustre (Fig. 11, A) and the inside surface was scoriaceous. The globules were found by the Dominion Laboratory to contain 95·6% of sulphur.

At the end of October 1954, the writer observed that black sulphur globules had been erupted from spring N85/6/73 at the Alum Cliffs. These globules were up to 9 in. in diameter, and of irregular shape (Fig. 11, B), but otherwise were similar to those from Whangioterangi. When first observed they looked very fresh, and it is thought that they were formed at the same time as those at Whangioterangi.

The globules were probably formed by a similar mechanism at both places. A concentration of sulphur may exist beneath the area, either disseminated in the country, or as an accumulation at the bottom of the respective lakes, or more likely as a deposit in fissures. Sudden reduction in pressure caused by earthquake fissuring could give rise to flashing of water at depth. The resulting steam would probably be of sufficiently high temperature to melt the sulphur. A mixture of molten sulphur, water vapour, and gas appears to have been injected into comparatively cool water, freezing the sulphur and forming a hard glazed skin (Fig. 12, B). The hollow internal cavity (Fig. 12, A) was probably originally filled with water vapour and gas. The difference in size of the globules



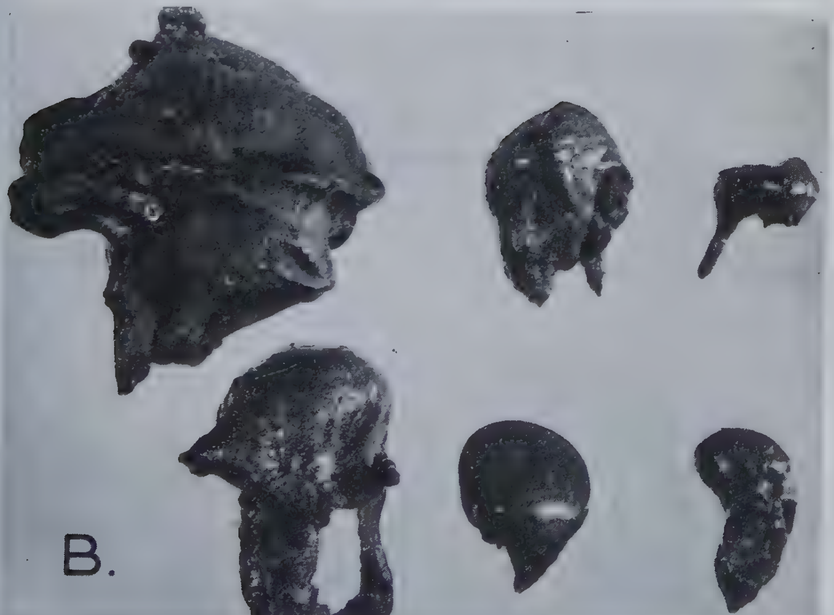


FIG. 11A.—Small sulphur globules formed at Whangioterangi, spring N85/6/85, during seismic disturbances in 1954. The grid represents ½ in. squares.

FIG. 11B.—A representative selection of the sulphur globules formed at spring N85/6/73. Note the "tail" which is a characteristic feature of the globules, and the "hat" (globule top left) which is a feature of many specimens. The surface farthest from the tail usually shows denting, as if due to partial collapse immeditely prior to solidification. This latter feature is shown on the two centre specimens. Grid represents ½ in. squares.

Photos: W. J. Wilson.

is probably related to the temperatures at depth and the rate at which the water vapour and gas was released. It is thought that a fairly rapid release of gas would be necessary to produce the contorted shapes and large size of the globules found about spring N85/6/73, whereas those found at Whangioterangi could have resulted from a comparatively mild disturbance. The globules from Whangioterangi were still being washed down the stream in 1957, and it was thought by the writer that they might still be forming, but none has definitely been observed from spring N85/6/73 since the 1954 earthquakes.

On the floors of the craters near the Alum Cliffs, the majority of springs are acid, with small water discharge and copious evolution of gas. On the banks of the stream draining Whangioterangi, between the lake and the Bridal Veil Falls, gas escapes and sulphur is deposited. There are many unusual mounds, not unlike mud volcanoes, but formed of small pisolites—up to ½ in.—of sulphur and mud. All the mounds appear to be extinct and it is not known how they were formed. Sulphur is deposited in the overflow channels of several small seeps of warm water but the remains of small sinter formations testify that alkaline conditions existed in the past.

Below the Bridal Veil Falls, on the flat, infilled floors of the craters, are several shallow lakelets of warm water. Small springs occur on the margin of these lakelets and some deposit a yellow sludge of sulphur. Spring N85/6/75 has a raised sinter lip, but its water is now acid and deposits sulphur. Nearby, spring N85/6/76 is a typical "frying-pan" area of shallow, violently bubbling pools of thin mud. At the west end of this crater there are several mud volcanoes, spring N85/6/79.

Springs N85/6/70, 71, 72, and 80 are the only chloride springs in this locality. Spring N85/6/70 is an active geyser with a long period of 12 hours or more between eruptions. The main vent is connected with another smaller opening 2 ft south of the geyser, through which overflow occurs prior to an eruption. The geyser ejects its water to a height of about 4 ft, completely emptying the basin. The remaining chloride springs in the locality occur at the foot of the crater walls and are insignificant.

LAKE NGAKORO

The northern and deepest part of Lake Ngakoro, N85/6/88 (Fig. 6) occupies a steep walled explosion crater. The southern part is shallow, and is floored with sinter deposited by thermal activity which has been drowned by a gradual rise in lake level due to the rising swamps between Lakes Ngakoro and Orotu. The temperature of the lake is about 25° C, and gas rises from many places on its floor. The water is coloured green by algae.

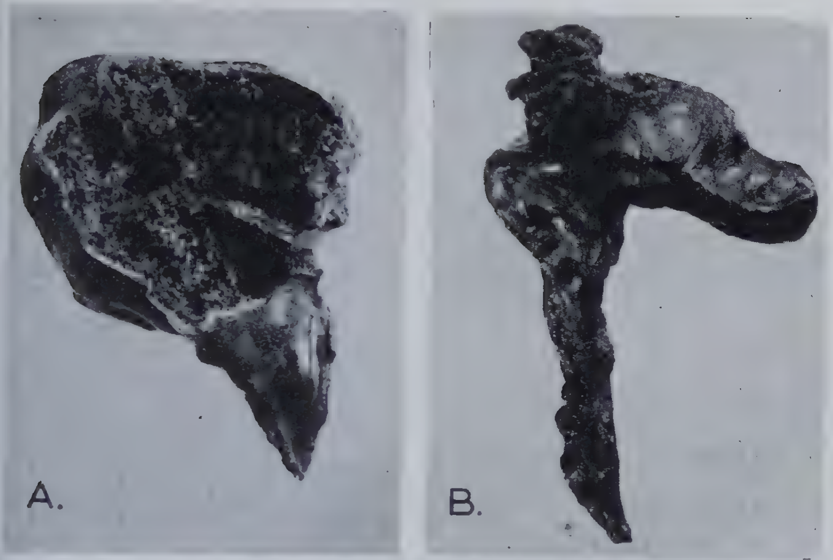


FIG. 12.—Sulphur globules from spring N85/6/73. A. Showing interior scoriaceous surface of globule. B. Showing detail of outer surface. Grid represents ½ in. squares. *Photos: W. J. Wilson.*

There are hot springs round the margins of the lake. The flowing springs are alkaline, with the exception of spring N85/6/86. The springs range in temperature between 80° C and 98·5° C, and all evolve considerable quantities of gas. The chloride waters deposit sinter, and prolific colonies of algae cover the terraces where the water has cooled below about 80° C. Springs N85/6/87 and 88 have built sinter cones to a height of about 4 ft, but now appear almost extinct.

About 20 chains south-east of Lake Ngakoro, on the edge of the swamp, are chloride springs, N85/6/142, 143, and 144. These springs are the southernmost of the Waiotapu activity. East of Lake Ngakoro, on the edge of the forest, is an area of warm ground containing several mud pots, springs N85/6/139, 140, and 141.

CHEMICAL COMPOSITION OF WAIOTAPU HOT SPRING WATERS

Hot spring waters were analysed for chloride, fluoride, bicarbonate, carbonate, and boric acid; pH and electrical conductivity were also measured.

Chloride was determined by the standard method of titration with silver nitrate solution, using potassium chromate as indicator.

Flouride was determined by the method of Milton *et al.* (1947), except that chrome azurol S was used as indicator. The method is suitable only for chloride waters low in sulphate, and when used for acid waters it gives unreliable results which can be corrected only if sulphate is also determined (see footnote, Table 1).

TABLE 1.—Some Constituents of Waiotapu Hot Spring Waters.

								Molecular ratios		
Spring No.	pH	Conductivity $\times 10^{-5}$ (Ω^{-1} cm^{-1})	Cl′ (ppm)	F′ (ppm)	HBO_2 (ppm)	HCO_3 (ppm)	H_2CO_3 (ppm)	Cl/F	Cl/ HBO_2	Cl/ HCO_3
N85/3/										
1	5·0	66	4·2	n.a.	Nil	Nil	Nil	9		
3	2·0	50	0·3	n.a.	Nil	Nil	Nil	0·01		
4	2·0	39	5·0	n.a.	Nil	Nil	Nil	0·2		
5	3·0	60	3·5	n.a.	Nil	Nil	Nil	3		
6	6·0	40·5	5·0	n.a.	Nil	Nil	Nil	4		
9	3·5	123	230·5	n.a.	9·8	0·4	Nil	169	29·0	1072
10	3·0	139	195·0	n.a.	9·4	Nil	Nil	20	26·0	
11	3·0	147	215·0	n.a.	14·5	Nil	Nil	22	18·0	
14	3·1	120	186·0	n.a.	8·7	Nil	Nil	21	26·0	
16	5·8	71	173·0	0·6	32·8	45·7	145·0	150·0	6·5	6·5
17	2·7	237	240·0	n.a.	12·0	Nil	Nil	169	24·0	
18	2·6	154	153·0	n.a.	7·0	Nil	Nil	30	27·0	
19	2·7	170	197·0	n.a.	8·7	Nil	Nil	24	28·0	
24	2·0	288	9·0	n.a.	Nil	Nil	Nil	0·5		
30	2·0	314	24·5	n.a.	Nil	Nil	Nil	3		
36	2·2	212	n.d.	n.a.	0·5	Nil	Nil			
48	2·7	74	0·7	n.a.	Nil	Nil	Nil	0·2		
52	2·4	191	141·0	n.a.	8·7	Nil	Nil	17	20·0	
56	2·4	186	93·0	n.a.	6·3	Nil	Nil	10	18·0	
58	3·5	31	4·0	n.a.	Nil	Nil	Nil	3		
59	3·0	45	3·0	n.a.	Nil	Nil	Nil	2		
61	3·0	53	3·0	n.a.	Nil	Nil	Nil	2		
63	3·2	23	4·7	n.a.	Nil	Nil	Nil	5		
64	3·4	26	3·0	n.a.	Nil	Nil	Nil	3		
67	3·4	31	4·6	n.a.	Nil	Nil	Nil	3		
68	3·4	31	4·6	n.a.	Nil	Nil	Nil	3		
69	3·1	33	3·0	n.a.	Nil	Nil	Nil	2		
70	3·2	30	2·0	n.a.	Nil	Nil	Nil	2		
72	3·2	30	2·7	n.a.	Nil	Nil	Nil	6		
75	2·2	195	2·7	n.a.	3·3	Nil	Nil		1·0	
97	2·2	206	4·5	n.a.	1·0	Nil	Nil	2	6·0	
100	3·0	19	4·0	n.a.	Nil	Nil	Nil	3		
101	5·2	52	4·0	n.a.	Nil	Nil	Nil	2		
102	7·4	23·3	4·7	n.a.	Nil	52·3	38·3	4		0·1
N85/6/										
3	2·8	81	6·0	n.a.	Nil	Nil	Nil	1		
4	1·9	533	8·5	n.a.	Nil	Nil	Nil	0·3		
4A	1·9	518	6·0	n.a.	Nil	Nil	Nil	0·3		
7	2·9	134	4·3	n.a.	Nil	Nil	Nil	1		
8	2·9	77	Nil	n.a.	Nil	Nil	Nil			
13	7·2	138	372	n.a.	n.a.	41	45	38		16
15	7·2	n.d.	493	n.d.	n.a.	90	10			10
20	8·6	223	660	8·0	25·2	64·7	1	45	31	17
29	7·2	235	636·5	n.a.	25·0	26	1·4	37	31	42·2
32	6·5	260	765	8·3	11·6	33	17	49	28	40
46	3·4	97	205	n.a.	10·0	Nil	Nil	68·7	25	
47	3·9	89	176	n.a.	Nil	Nil	Nil	85·8		
59	6·5	322	1040	6·9	3·8	39·5	21	81	33·5	45·3
50	8·2	333	1050	4·9	3·4	47	0·5	115	37·3	38·2

TABLE 1.—*Continued.*

Spring No.	pH	Conductivity $\times 10^{-5}$ (Ω^{-1} cm^{-1})	Cl′ (ppm)	F′ (ppm)	HBO_2 (ppm)	HCO_3 (ppm)	H_2CO_3 (ppm)	Molecular ratios Cl/F	Cl/HBO_2	Cl/HCO_3
51	8·2	348	1060	7·3	4·7	61	Nil	85	30	33
54	6·7	90	147	n.a.	6·2	104	32		29	2
64	6·4	n.d.	1880	5·5	110	230	136	180	21	14
86	3·5	228	593	2·9	22	Nil	Nil	110	33	
87	8·6	n.d.	733	n.d.	n.d.	173	1			7
88	2·5	317	585	4·6	24·2	Nil	Nil	68	30	
89	2·5	160	17	4·8	Nil	Nil	Nil	2		
95	4·0	63	10	2·4	Nil	Tr	83	2		
96	3·6	178	468	3·6	24·8	Tr	92	70	23	
97	6·3	458	1778	6·6	80·6	97·5	80	144	27·5	31·3
99	7·4	452	1604	6·7	73·4	118	8	128	27	23·4
100	6·7	147	244	3·1	Nil	745	252	42		0·6
101	6·7	205	478	3·4	10·4	645	172	75	57	1·3
102	7·2	138	241	3·2	Nil	730	83	40·4		0·6
103	7·6	287	852	8·7	25	298	12	52·5	42	0·5

nd = not determined   na = result not accurate due to SO_4''

NOTE.—With the method of analysis used, F′ is not accurate when SO_4'' is greater than 50 ppm. F′ has been omitted where SO_4'' is suspected, but since no SO_4'' determinations were made, all F′ determinations should be treated with caution.

Bicarbonate and carbonate were determined as follows: The pH was measured using a pH meter and glass electrode system on 100 ml. of the sample. The pH was then adjusted to 8·3 by the addition of 0·1N acid or alkali and the carbon dioxide (together with hydrogen sulphide) determined by titration with 0·1N hydrochloric acid to pH 3·7. As the method includes any hydrogen sulphide present with the carbon dioxide, hydrogen sulphide was determined by titration with 0·01N iodine solution using starch as indicator, and subtracted from the result of the first titration. The proportion of total carbon dioxide present as bicarbonate and carbonate was read off from a graph, prepared by calculating from the dissociation constant at 15° C, the molar proportions at various pH values.

Boric acid was determined by direct titration with the addition of mannitol, using the method described by the American Public Health Association (1955, p. 46), except that a glass electrode was used.

During the early stages of the survey all springs were sampled, but later sampling was confined to the flowing springs, since it was found that the perched acid-sulphate waters (which are in the majority at Waiotapu) were so weakly mineralized that their analyses proved of little use. The analyses are presented in Table 1.

The spring waters have been divided into four classes, acid-sulphate waters, sulphate-chloride waters, chloride waters, and bicarbonate-chloride waters.

Acid-sulphate Waters

These waters have pH of 2-5, acidification being due to superficial oxidation of hydrogen sulphide. Sulphate is the chief anion, and chloride is absent or is present as a trace only. Temperatures vary from 30° C to 99° C; discharge of water is small or absent and evolution of gas is considerable. Springs of acid-sulphate water predominate numerically throughout the whole Waiotapu thermal area, mainly at situations above the water table, but commonly they occur associated with the other types of water. They are believed to derive heat from steam boiling off at depth from chloride ground water and condensing again near the surface. Most depend on rainfall for their water supply and during the summer months may form "frying-pan" areas, or dry up altogether resulting in steam jets.

The springs along Ngapouri Fault, on the lower eastern flanks of Maungaongaonga, form a special group of acid-sulphate waters, in that many have strong discharges during the winter months, but during periods of low rainfall overflow may cease altogether. Most of the springs have a uniform temperature of about 50° C and are weakly but uniformly mineralized. This suggests that steam, boiling off from the main body of ground water at depth, is condensed by perched ground water supplying the springs. Further evidence for steam heating is the low chloride-fluoride ratios of these waters (Wilson, 1953; Ellis and Wilson, 1955). Wilson (pers. comm.) has pointed out that steam at low pressure carries some ammonia (not analysed for in this survey), boric acid, and fluoride, but no chloride. Low values for the chloride-ammonium, chloride-boric acid, and chloride-fluoride ratios indicate that the water has scrubbed these constituents from steam. Springs N85/3/3, 4, and 5, in the crater on the western flanks of Maungakakaramea also have low chloride-fluoride ratios and contain acid-sulphate water.

The chemical analyses do not give any indication of the type of water at depth, but it is most likely chloride water, similar to that under the main thermal area.

Sulphate-chloride Waters

The springs in this group differ from the acid-sulphate waters in that they contain approximately 200 p.p.m. of chloride ion and have moderate discharges (1-3 litres per second) which are little affected by rainfall. The water is acid, pH 2·5, and temperatures are near boiling. In these respects these waters are similar to those at Waiora Valley, Wairakei. At Waiotapu, sulphate-chloride waters are restricted to springs on the north-eastern shore of Rotowhero and on both banks of Hakerateke Stream where it skirts the southern base of Maungakakaramea.

Sulphate in thermal waters is generally considered to be due to oxidation of hydrogen sulphide, and ideal conditions for this are, (1) reasonably high concentration of hydrogen sulphide, (2) little or no discharge, because of a perched position, hence inadequate flushing of the spring system, and (3) porous near-surface strata allowing aeration.

The springs under consideration emerge at the water table and have a considerable discharge, hence it is unlikely that any dissolved hydrogen sulphide would have prolonged contact with oxygen. The waters contain iron, suggesting that they have attacked the country rock, and on this basis it is thought either that acidification occurs at considerable depth, or that the waters are re-circulated after near surface oxidation of hydrogen sulphide to sulphate ion has occurred. Both these mechanisms provide a means whereby water, high in sulphate, can have prolonged contact with wall rocks.

Mr S. H. Wilson (pers. comm.) considers it unlikely, owing to the supply of oxygen being inadequate, that hydrogen sulphide would oxidize in any quantity to sulphuric acid at depth; the oxygen in ground water is only sufficient to provide about 12 p.p.m. of sulphate ion. He prefers the concept of re-circulation, and suggests that in areas of steaming ground (e.g. Maungakakaramea) hydrogen sulphide is partly oxidized to sulphate ion near the surface, but is not volatile and will be washed down and will accumulate at lower levels. Most of it will attack the rocks and give sulphates. There will be an equilibrium underground, perhaps neutral or very slightly acid, but when the water reaches the surface, more hydrogen sulphide may be oxidized, or iron hydroxide precipitated, providing enough hydrogen ions to account for the low pH.

Another simple explanation for these waters is that they result from the mixing of chloride water with an acid sulphate water. Boiling water containing 1,800 p.p.m. of chloride ion occurs at the south end of the thermal area, and if it is assumed that this has been diluted by cold ground water at 13° C, containing no chloride, the temperature of the mixture to give a water containing 200 p.p.m. of chloride ion would be approximately 23° C. To account for the near boiling temperatures in the sulphate-chloride springs, the diluted chloride water would need to be mixed with a sulphate water at a temperature higher than boiling.

This immediately suggests a steam feed as the source of the sulphate, and suggests that it may be due to deep condensation of volcanic emanations containing oxidized sulphur and possibly alkali halides. Considering that the nearby crater, now occupied by Lake Ngahewa has erupted within the last 900 years (for radiocarbon dating—see later section) this is not so unlikely as it might first appear.

Chloride Waters

The chloride waters contain chloride as the chief anion, up to 1,800 p.p.m., and bicarbonate and sulphate are present in small quantities. The waters are near neutral, ranging in pH from 6 to 8·5, alkalinity being indicated by the presence of bicarbonate. Temperatures range between 60° C and 101° C (Postmistress, spring N85/6/20), and generally the springs with highest temperatures have the highest chloride content. Most evolve a moderate quantity of gas, mainly carbon dioxide, but from some springs the odour of hydrogen sulphide is perceptible. The chloride waters deposit a greyish sinter, and remnants of old sinter formations testify to more widespread surface discharge of such water in the past. The chloride waters occur within an area extending from the base of Maungakakaramea south to the tourist reserve, and occur close to the streams where the main springs would be expected. A considerable amount of chloride water seeps into the streams. These waters are similar to those at Geyser Valley, Wairakei, but contain less chloride.

Fig. 13A shows iso-chloride lines over the Waiotapu thermal area. The east and west boundaries are approximate, owing to the absence of hot springs. The chloride maximum occurs in the vicinity of Champagne Pool, extending northwards along the course of the Waiotapu Stream to Weir's Road, and on the south-east shore of Lake Ngakoro. It is thought likely that the chloride high is continuous at depth between Champagne Pool and Lake Ngakoro. Chloride water apparently finds access to the surface through shattered rock associated with craters and faults in this area, but some escapes into the country and appears as springs some distance away, its dilution by ground water increasing with the distance from the craters. A plot of chloride/fluoride ratio (Fig. 13C) produces a similar pattern to the iso-chloride lines; the implications of this are discussed in a later section.

Bicarbonate-chloride Water

South from the chloride area there is a gradual change of water, with bicarbonate content increasing to approximately 500 p.p.m., and chloride up to 1,800 p.p.m. These constituents reach highest concentrations in the springs on the south-eastern shore of Lake Ngakoro. The springs range in temperature from 70° C to 99° C, and all evolve considerable quantities of gas, mainly carbon dioxide, but hydrogen sulphide is detectable in some areas.

The springs with large overflows generally have high chloride content and moderate bicarbonate content, whereas the springs with small discharges have moderate chloride and high bicarbonate. Both exist side by side. It is apparent from surface observations that the area about Lake Ngakoro is the most gas-rich locality at Waiotapu, and it is likely that the high concentration of bicarbonate and gas are related. Waters with bicarbonate and with some chloride are of shallow origin and are formed by attack by carbon dioxide on the rock minerals. It is likely that chloride water exists below the zone of bicarbonate-chloride water.

HEAT OUTPUT AT WAIOTAPU

Measurements of the natural heat flow of the Waiotapu thermal area were made mainly by Mr R. F. Benseman, Dominion Physical Laboratory, and the writer is indebted for the use of Mr Benseman's unpublished figures for this section. The writer measured the heat flow from the Maungaongaonga fumarolic area, and the springs in the craters to the north-east as far as Maungakakaramea.

Temperatures in hot springs were measured with a standard maximum thermometer, or thermocouple and potentiometer. Fumarole temperatures were measured with the thermocouple and potentiometer. The measurements of discharge from springs on Ngapouri Fault were made using a 90° V-notch after a long period of excessive rainfall when the spring discharges were a maximum. Individual spring flows were not measured over, the remainder of the thermal area, where most of the hot water enters the streams as seepage, but the heat flow as water from the area was determined by measuring the increase in flow and temperature of the streams as they passed through the hot area. Allowances were made for cooling along the course of the streams. Steaming ground at Maungaongaonga was measured using a calorimeter (designed by R. F. Benseman), and the figures were adopted as the basis for estimation of heat flow on Maungakakaramea. Other areas of steaming ground were estimated using the growth of shrub as an indication of the intensity of the activity. The velocities of discharge of fumaroles at Maungaongaonga were measured with a pitot tube.

The results are summarized in Table 2.

TABLE 2.—Natural Heat Output of the Waiotapu Thermal Area. (Kcal/sec)

Source of Heat	Main Thermal Area*	Maunga-ongaonga	Maunga-kakaramea	Totals
Discharge from Springs Evaporation from Pools Evaporation from Springs	148,200	1,800	16,000	166,000
Steaming Ground	55,400	26,900	35,400	117,700
Conduction†	12,000	1,900	4,004	18,300
TOTALS:	215,600	30,600	55,800	302,000

†Conduction factor calculated on basis of k (ground conductivity) $= 3 \times 10^{-3}$cgs units

*After Benseman.

WAIRAKEI AND WAIOTAPU HOT SPRINGS—A COMPARISON

There is one very striking difference between Wairakei and Waiotapu, the two thermal areas so far studied in detail. At Wairakei the hot springs are concentrated within several comparatively small areas, Geyser Valley, Waiora Valley, etc., but at Waiotapu, with the exception of the Maungaongaonga fumarolic area, there is no comparable concentration of thermal activity. Few hot springs at Waiotapu have overflows exceed 5 litres per seond, and most activity is confined to extensive, but not intensive areas of steaming ground, to hot water seepages along the banks of streams, and to large sulphate pools which discharge a mere trickle and lose most of their heat by evaporation.

The difference between Wairakei and Waiotapu activity is probably due to several causes. Geyser Valley is deeply eroded, and in the vicinity of the thermal area where the water table stands higher than in surrounding cold area, the stream intersects the water table, thus inducing a concentrated flow of hot water. Waiotapu is not eroded as deeply as Geyser Valley, and is more like Waiora Valley, though the water discharge from individual springs is less at Waiotapu than at Waiora Valley.

At Wairakei, impermeable siltstones and mudstones of the Huka Formation (Steiner, 1953) cap an extensive aquifer, and beneath the siltstones and mudstones pressures are often higher than hydrostatic head. Chloride water escapes to the surface through fissures in the mudstone. Analyses of Champagne Pool, Geyser Valley (spring N94/4/97) show that its water is similar to that tapped by drillholes at depth, indicating that the spring is directly connected with the aquifer because its water has not greatly been contaminated by cold ground water. The same is probably true of other major springs in Geyser Valley. It seems likely that in areas of concentrated thermal activity, the feeding fissures up which the hot water migrates will approach or even reach the surface. Hence, the pressures in the aquifer and the permeability of the country through which the hot water migrates to reach the surface, will affect the intensity of the natural thermal activity.

At Waiotapu, comparable sub-surface geological conditions exist. Impermeable siltstones and mudstones occur at shallow depth—from near the surface down to about 150 ft—and cap an aquifer. Since most of the heat is dissipated over a large area without strong concentrations at individual points, except along the Ngapouri Fault, the inference is that the siltstones and mudstones do not form so effective a cap as the Huka mudstones at Wairakei. At Waiotapu pressures are relatively lower beneath them, and discharges are accordingly lower. The most thermally active localities are those adjacent to hydrothermal explosion craters, where the hot water presumably rises through shattered zones.

SOURCE OF WATER

Fig. 13A shows iso-chloride lines drawn from all available springs that have sufficient overflow for the water to be considered representative of that in the supplying aquifer. Maximum chloride concentrations occur at Champagne Pool and springs N85/6/97 and 99 at Lake Ngakoro, and these discharge chloride water similar chemically to that tapped by drillholes at Wairakei (Table 3). The iso-chloride lines show that water high in chloride finds direct access to the surface at springs in the Champagne Pool-Ngakoro area, and suggest that most of this water escapes into the country in permeable beds beneath the mudstones, where it is progressively diluted and finally reaches the surface through faults and joints in the mudstone. The diluted chloride water only has been tapped by drillholes down to 1,500 ft in the area away from the suggested source.

TABLE 3.—Comparison of Wairakei and Waiotapu Thermal Waters.

Source	Molecular Ratios			
	Cl/F	Cl/B	Cl/As	Cl/S
Wairakei*				
Bore 4/2	143	23	980	144
Bore 12	128	24	950	120
Bore 15	125	22	1170	112
Bore 20	127	24	1060	100
Champagne Pool (N94/4/97)	143	24·5	1020	160
Waiotapu				
Champagne Pool (N85/6/64)	180	21		
Spring N85/6/97	144	27		
Spring N85/6/99	128	27		

*From Ellis and Wilson (1955).

Contours of chloride-boric acid ratio shown in Fig. 13B indicate that this ratio is approximately constant over the whole area, except for two springs at the southern corner of Lake Ngakoro which are low in boron and give higher values for the ratio. This general picture is consistent with the conclusions from the iso-chloride lines.

The chloride-fluoride ratio as shown in Fig. 13C, however, is not constant, but shows a gradual decrease away from the Champagne Pool-Ngakoro area. Ellis and Wilson (1955) have shown that at Wairakei the ratio of chloride to fluoride and to boron, arsenic and total sulphur in the complete drillhole discharges, and in the chloride water of the natural activity, are nearly constant, and use this as evidence for a homogeneous aqueous solution at depth. Wilson (pers. comm.) considers that the ratios are constant in any thermal area supplied from a single source, and that water from different sources will have different values for the ratios. The values for the chloride-fluoride ratios at Waiotapu appear to contradict this.

The similarity in pattern of the iso-chloride lines and the contours of chloride-fluoride ratio suggest that the factor governing the progressive decrease in total chloride in spring waters away from the Champagne Pool-Ngakoro area is also the primary cause of its decrease relative to fluoride. In other words, the dilution process involves also a relative increase in fluoride, which would indicate that the ground water diluting the high chloride water in the upper aquifer is relatively high in fluoride.

As mentioned previously, the waters of springs on Ngapouri Fault are steam heated, and this has resulted in relatively high fluoride content (analyses springs N85/3/58-72, Table 1). It is suggested that a ground water contribution from this direction is responsible for the decrease in chloride-fluoride ratio described above. Even the water of Lake Ngahewa, spring N85/3/6, which contains no known thermal contribution has relatively high fluoride.

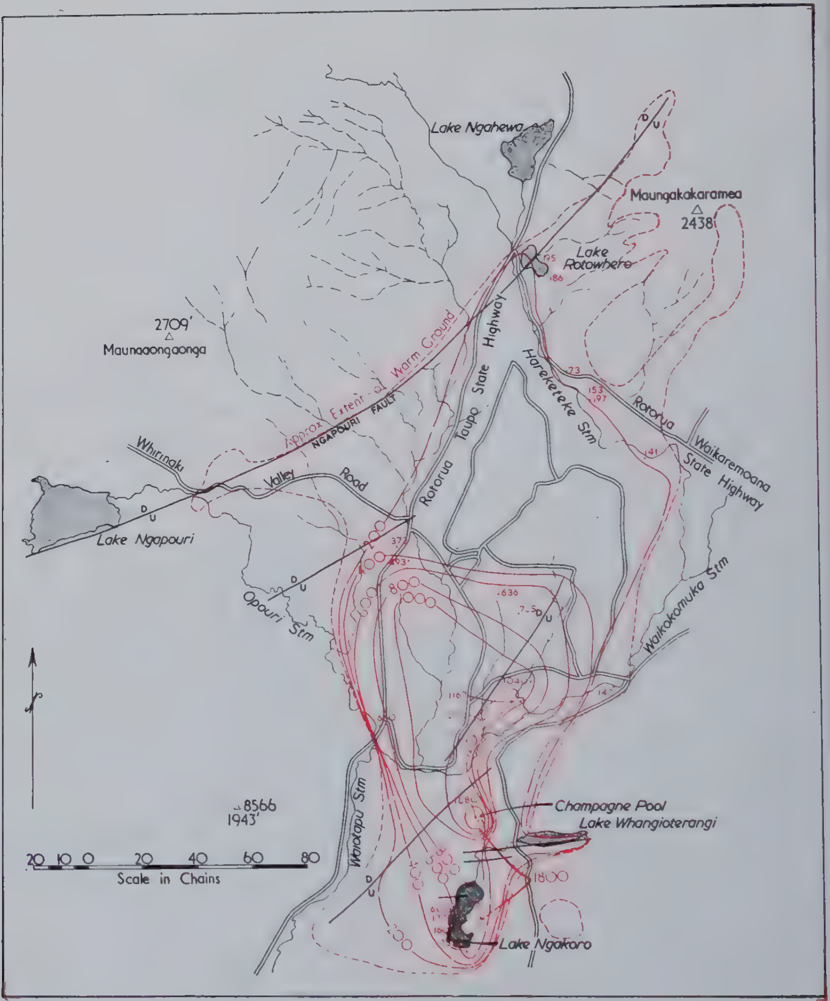
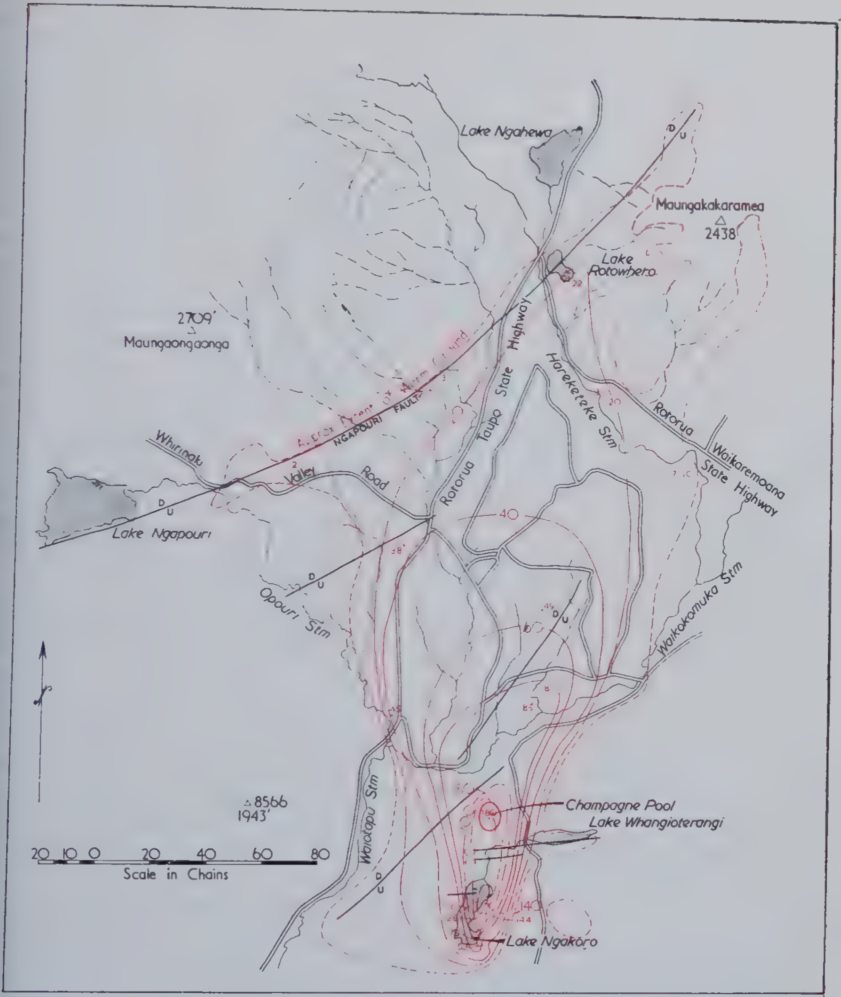


FIG. 13A.—Iso-chloride lines, drawn from hot spring analyses.

Hence it is concluded that there is only one source for the hot spring waters at Waiotapu. This is an aquifer containing high chloride water at depth greater than 1,500 ft which probably underlies the whole area and finds access to the surface at springs in the Champagne Pool-Ngakoro area. The chemical characteristics of the majority of the hot spring waters result from mixing of this high chloride water with a perched, steam-heated ground water, relatively high in fluoride, that migrates down from the Ngapouri Fault area on the slopes of Maungaongaonga. It is suggested that the steam in this area may result from flashing of high chloride water at depth, implying that temperatures beneath the Maungaongaonga region are very high. Further implications of this are discussed in a later section.



~~FIG. 13B.—Contours of chloride-boric acid ratio, drawn from hot spring analyses~~

FIG. 13C.—Contours of chloride-fluoride ratio, drawn from hot spring analyses.

RECENT HYDROTHERMAL ERUPTIONS AND THEIR EFFECT ON THE HYDROTHERMAL SYSTEM

The Waiotapu landscape is pitted with many small craters, the sites of brief, but violent hydrothermal outbreaks of the steam-blast type. Thomas (1888, p. 9) and Martin (1895, p. 88) mention the existence of crater-shaped hollows on the sides of Maungakakaramea and farther south in the Waiotapu Valley, and suggest they were formed by hydrothermal explosion. Grange (1937, p. 37) stated that Lake Okaro occupies an explosion crater. More recently, Cross (unpublished Soil Bureau

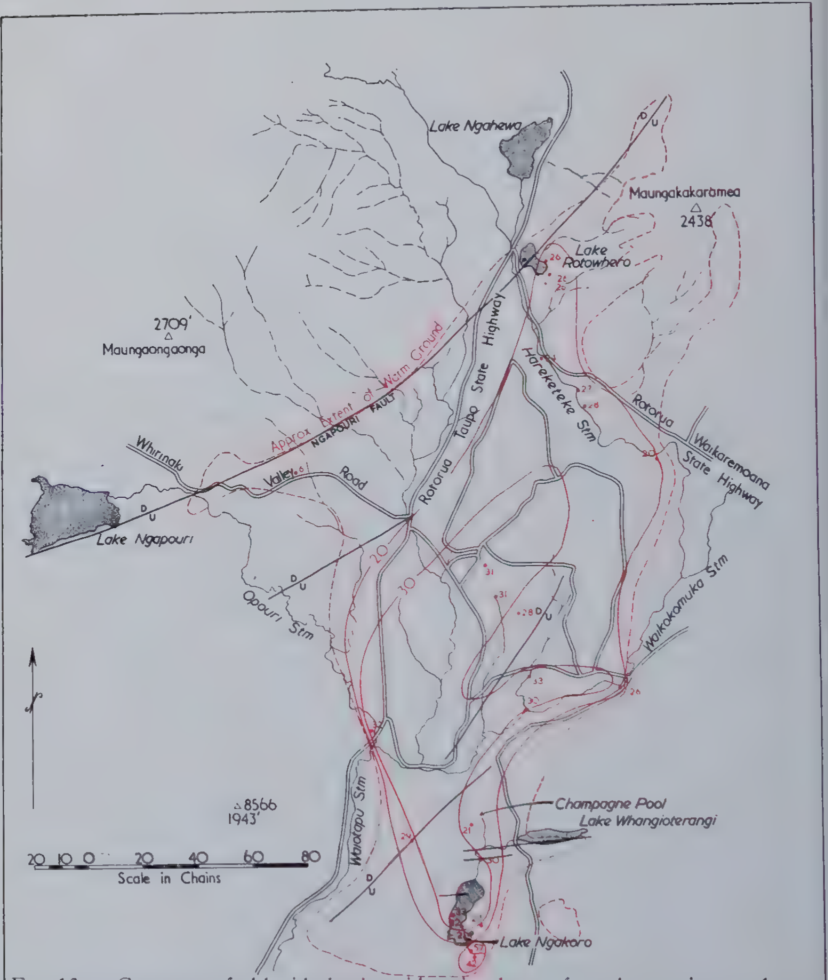


FIG. 13B.—Contours of chloride-boric acid ratio, drawn from hot spring analyses.

~~FIG. 13c.—Contours of chloride-fluoride ratio, drawn from hot spring analyses.~~

Report, pers. comm.), describes the craters along and on the downthrown side of Ngapouri Fault. There is a further group of craters near the south of the Waiotapu thermal area. The craters described by Cross will be briefly mentioned in this paper but those farther south need more detailed treatment because of their close association with present-day thermal activity. The eruptions are classed as hydrothermal because only hydrothermally altered country rock was ejected.

Craters on Ngapouri Fault

Of the five major craters on Ngapouri Fault, three are now occupied by lakes (Fig. 2). Lakes Ngapouri and Opal are cold, and Lake Rotowhero is heated by hot springs (N85/3/8-13) on the south-eastern shore, and by springs emerging in the lake bed. The remaining two craters are on the flanks of Maungakakaramea. One north-east of Rotowhero has springs (N85/3/3, 4, and 5) on its floor, and steam issues from its walls. The other, south-west of Opal Lake, has a swampy floor and appears cold. Between lakes Ngapouri and Rotowhero there is a line of small craters containing hot springs which are relatively minor foci of eruption, but indicate that Ngapouri Fault is continuous at depth between Lake Ngapouri and Opal Lake. Lake Ngahewa is cold, and occupies a crater on the downthrown side of Ngapouri Fault. Lake Tutaeinanga is approximately one mile west of Lake Ngapouri and is cold, as is Lake Okaro, between Maungakakaramea and Waimangu. The relationship between these two craters and those which are defintely associated with Ngapouri Fault is not clear.

There are other craters on Maungakakaramea whose association with Ngapouri Fault is not obvious. One, high on the western slopes of the volcano contains a mature forest, and a 60 ft section of ejecta exposed on the northern rim indicates that this crater has been the source of most of the material covering the northern slopes of Maungakakaramea, though much of this has subsequently been washed off the steep slopes. On the eastern slopes there is another conspicuous crater which has ejected a small quantity of hydrothermally altered country rock. Between this crater and the Opal Lake crater are several minor craters, aligned parallel to the Ngapouri Fault (Fig. 2).

Present-day activity along Ngapouri Fault extends from the Maungaongaonga fumarolic area to the warm western slopes of Maungakakaramea. Within the past 100 or so years, activity extended farther to the north-east to about the Opal Lake, where crystalline sulphur, several inches thick, is overlain by Rotomahana Mud, indicating that there was activity there shortly before the eruption of Tarawera and Rotomahana in 1886. Hochstetter (1864, p. 144) who viewed Maungakakaramea from Rotomahana in 1853 stated that the mountain was steaming "from foot to summit . . . so that the mountain was quite enveloped in steam as if it were a volcano." Thomas (1888, p. 9) stated that "in the early morning the steam can be seen rising in thick clouds from the mountain giving an impression of a formidable degree of activity." Another early description (Anon. 1891) reads "On its face are large tracts of variegated clays, presenting all the colours of the rainbow and a great many more besides, this effect having been produced by hydrothermal action. It looks as if the mountain has been fried and grilled, baked and scorched in turn, till it is decidedly 'overdone'. As we drove past it [Maungakakaramea] we saw heavy clouds of steam rising from it." The author has recently viewed the northern slopes of Maungakakaramea on a frosty morning from Waimangu, which is in approximately the same direction from the mountain as Rotomahana where Hochstetter made his observations, but no steam could be seen. This suggests that activity on the northern slopes of Maungakakaramea has declined noticeably in the last 60 years.

At the south-western end of Ngapouri Fault there is also evidence that activity has been more widespread. On the Waikite Valley Road, near the Maungaongaonga fumarolic area, hydrothermally altered ash showers are exposed where the surface is now cold, and thermal activity has been more intense at the small barren areas south-west of Lake Ngapouri.

Ejecta from craters near Maungakakaramea is more strongly altered than that from Okaro and Ngapouri. This suggests that the rock near Maungakakaramea had either been subject to hydrothermal alteration for a longer time than the rock farther north-east and south-west, or that temperatures were higher near Maungakakaramea. Whatever the cause, the indication is that the hottest ground was near Maungakakaramea. This is supported by the present distribution of heat flow along Ngapouri Fault.

If all the eruptions occurred from a common hydrothermal system, this system must have been extensive, since evidence of pre-eruption thermal activity can be found over an area of eight square miles. The implication is that for such a large block of country to heat up, heat losses at the surface must initially have been low compared with the rate at which heat was being supplied from below. The system would eventually reach equilibrium, when the heat gain would equal the heat lost from the system by surface thermal activity.

After the eruptions, shattered zones beneath the craters would channel the hot water to the surface, resulting in accelerated heat output and allowing cold ground water to flow in from the margins. Permanent shrinkage of the hydrothermal system would result, provided the shattered zones remained open to the passage of hot water. Possibly, flashing is still occurring in the Maungaongaonga area and the hydrothermal system may still be shrinking, explaining the apparent recent cooling at Maungakakaramea, or the cooling could be due to a dimunition in the heat supply from the source. As pressures dropped in the hydrothermal system so would the surface thermal activity correspondingly decrease, and on the high ground traversed by Ngapouri Fault where the water table is low, the springs might cease to overflow, though outward migration of hot water underground could still continue.

Craters at the South of the Waiotapu Thermal Area

The craters to the south (Fig. 2) are on lines trending at 080° which are interpreted as faults. Many of the craters are elongated in the same direction, and Lakes Ngakoro and Whangioterangi occupy the largest of these, and other smaller craters form the Alum Cliffs. Champagne Pool occupies a crater, and many of the smaller pits throughout the southern part of the thermal area have had an explosive origin. All these craters contain thermal activity, described in an earlier section.

The rock ejected from the southern group of craters is less altered than that from the craters near Maungakakaramea, suggesting that thermal activity in the southern area prior to the eruptions was less intense, or more recent, than that near Maungakakaramea. A great variety of rock is found in the Ngakoro deposit. The lenticular ignimbrite found in the drilled rock sequence between approximately 800 ft and 1,900 ft (Mr J. Healy, pers. comm.), is common. Unaltered rhyolite and dacite are also common, the latter as rounded stones, apparently water transported. Mr A. Steiner (pers. comm.) has reported that this rock greatly resembles the dacite of Maungaongaonga. It appears that the eruptive source was a deep aquifer which was effectively capped from the overlying rocks, resulting in the near surface rocks not being greatly altered. The source of the hot water to this aquifer may have been to the north, near Maungakakaramea, because of the higher rank alteration of ejecta from craters in this locality.

AGE OF THE ERUPTIONS

Tree remains have been discovered beneath the deposits from Okaro, Ngahewa, the upper crater on the western slopes of Maungakakaramea, and Lake Tutaeinanga. Carbonized fragments having the appearance of small branches have been collected from the contact between the breccia from Lake Ngakoro and the Taupo topsoils, and await dating. At the time of writing, radiocarbon datings of the Okaro and Ngahewa eruptions have been received (Fergusson & Rafter, 1959).

The sample from Lake Okaro (N.Z. ^{14}C No. 166) was projecting from the bank on the southern shore of the lake 2 ft above the contact between the Okaro breccia and the Taupo topsoils. It was identified by Forestry Research Institute, Whakarewarewa, as a 70-year-old Manuka (*Leptospermum* spp.), and had the bark intact. Age: 840 ± 50 years.

The sample from Lake Ngahewa (N.Z. ^{14}C No. 169) was collected by the author from a branch identified by Forestry Research Institute as Matai (*Podocarpus spicatus*), exposed in a gully entering the north-west side of the Ngahewa crater. The gully is eroded through the Ngahewa breccia into the ash showers beneath, and at the contact many tree trunks are exposed lying in a direction away from Lake Ngahewa, as if felled by a blast radiating from the crater, or by a directional blast from one of the craters on Maungakakaramea. The sample was secured from the outer ten growth rings of a branch with bark intact. Age 900 ± 40 years.

It can be seen from Table 4 that the radiocarbon dating method gives an overlay for the Ngahewa and Okaro eruptions of between 860 and 890 years ago. Considering that the fossil Taupo soils beneath the young ejecta show a similar stage of development over the whole area affected by the hydrothermal eruptions, and that the craters are aligned lineally, it is thought likely that the eruptions occurred either simultaneously, or within a brief time interval between 860-890 years B.P.

TABLE 4.—Radiocarbon Ages.

Sample	To Date	^{14}C Age (years B.P.)	Limits (years B.P.)
N.Z. ^{14}C 10	Kaharoa Eruption	930 ± 70	860 — 1000
N.Z. ^{14}C 169	Ngahewa Eruption	900 ± 40	860 — 940
N.Z. ^{14}C 166	Okaro Eruption	840 ± 50	790 — 890

MECHANISMS FOR TRIGGERING THE ERUPTIONS

Possible mechanisms for triggering the eruptions are:

(1) Accumulation of pressures within the hydrothermal system until they exceeded lithostatic load and disrupted the overlying rocks .

(2) Intrusion of magma at depth, upsetting the equilibrium that existed in the hydrothermal system because of intensified thermal effects.

(3) Tectonic movements allowing hot water to rise to levels where its temperature exceeded boiling point for depth conditions.

The Kaharoa pumice eruption (Grange, 1937), which the writer considers preceded the building of Tarawera dome, the southernmost of three coalescing rhyolite domes, Wahanga, Rauwahia, and Tarawera, often referred to collectively as Mount Tarawera, has been dated by the radiocarbon method (Fergusson and Rafter, 1955, 1957). Its age, and those for Ngahewa and Okaro eruptions are tabulated in Table 4, which shows that the Kaharoa eruption also overlaps the Waiotapu eruptions between 860-890 years B.P. It is therefore probable that all these eruptions were simultaneous. One possible inference is that the hydrothermal eruptions at Waiotapu were initiated by the more violent volcanic disturbances at the site of Tarawarea, and if this were so, a gradual build-up within the hydrothermal system until pressures exceeded lithostatic load can be ruled out as the immediate cause of the eruptions.

Perhaps the most valuable evidence in favour of the intrusion theory is the minor high gravity anomalies (Mr F. E. Studt, pers. comm.) due to the presence of dense rock beneath the areas in which the craters occur. Even assuming these are due to intrusion, there is no means of determining the time relationship between the intrusive event and the hydrothermal eruptions. Impermeable beds beneath which pressure could accumulate are not an essential part of the intrusive theory, since the balance existing in the hydrothermal system could be upset by sudden intense heat flow from the intrusions. On the other hand, if intrusion did trigger the eruptions it might be expected, though it is not essential, that some new lava would be ejected, but as far as is known only hydrothermally altered country rock was erupted. During the 1886 eruption of Mount Tarawera, basaltic lapilli were erupted from the Black Crater (Smith, 1887, p. 61), whereas nearby Inferno, Echo, and Southern Craters appear to have erupted only hydrothermally altered country rock. This is difficult to explain, and suggests the possibility that the craters at Waiotapu may have been potential ejectors of new magma, as presumably were Inferno, Echo, and Southern Craters. Magma is most likely to be intruded along existing fault planes, which could explain the alignment of the craters, but this can be explained just as satisfactorily by the tectonic theory.

Undoubtedly, local earthquake shocks would have accompanied the Kaharoa Eruption, and these could have produced movement on faults in the Waiotapu district, such as were observed along the southern continuation of the 1886 fissure (Smith, 1887, p. 31) following the Tarawera Eruption. The opening up of existing faults, and perhaps the formation of new ones would allow hot water to rise from depth. Similar events took place at Rotorua during the eruption of Tarawera and Rotomahana. Smith (1887, pp. 38-39) stated, "At Rotorua, the site of innumerable hot springs . . . a very great increase in action was noticed on the night of the eruption. . . . The changes, indeed, are so remarkable that it is a matter of surprise that they should have been doubted." As mentioned previously, the pre-eruption hydrothermal system at Waiotapu was extensive and may have been expanding, and this together with the intensely altered nature of the ejecta from some craters, suggests that high temperatures probably existed at depth. If faults or joints were to open and allow this hot water to rise to a level where its temperature exceeded boiling point for depth conditions, eruptions could result. The disturbances described in an earlier section, at Whangioterangi and spring N85/6/73, and attributed to seismic disturbances during 1954 were probably of a similar, but milder nature.

Both these hypotheses account for the observed facts, and no definite distinction seems possible unless further facts come to light which help to distinguish between the cause and the effect. Of the two, the writer prefers the tectonic theory.

ACKNOWLEDGEMENTS

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SOME OBSERVATIONS ON THE RADIOACTIVITY OF NGAURUHOE AND WHITE ISLAND LAVAS

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Summary

Radioactivity measurements were made on Ngauruhoe lavas of various age, and also on xenoliths found in the Ngauruhoe flow of September 1954. The variation in α count rates of the Ngauruhoe lavas erupted in 1954 is due to the presence of a varying amount of small feldspathic xenoliths or their absence. The radioactivity is roughly proportional to the potash content.

A radioactivity survey with a geiger tube was made on the White Island volcano. Here the more recent lavas exhibit a higher radioactivity than the older lavas.

INTRODUCTION

Ngauruhoe and White Island are active volcanoes situated at the opposite ends of the Taupo Volcanic Zone in the North Island of New Zealand. Ngauruhoe, characterized by lavas of basaltic andesite, is a young cone on the flank of the dormant composite volcano Mt. Tongariro. Andesite lavas only were erupted from White Island.

During the 1954 eruptions of Ngauruhoe (Gregg, 1956) the opportunity was taken to investigate the radioactivity of the old and new lavas and of some xenoliths of gneissic origin found in the new flows. The Ngauruhoe lava erupted in 1954 is an olivine-bearing basaltic andesite, and the older lavas from Ngauruhoe and Tongariro vary from basalts to basaltic andesites.

Ferretti-Sporzini (1954) investigated lavas from Vesuvius and found that the specific radioactivity is greater the more recent the lava. This suggests that an enrichment process is taking place in the Vesuvian magma chamber. Gibbs and McCallum (1955) using geiger tubes, found similar but less marked trends in some New Zealand ash showers. Their work initiated a radioactivity survey of White Island lavas *in situ* in March 1956. Several andesite lava flows are present on White Island (Kaplan, 1956) and eruptions may have commenced in the late Pleistocene (J. Healy, pers .comm.).

EXPERIMENTAL DETAILS

To ensure fresh and unleached samples of the new Ngauruhoe lava, several specimens of the September 1954 lava flow and its xenoliths

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were collected immediately after the eruption. In addition, samples of older lavas of known age were collected from Ngauruhoe and Tongariro, so that their radioactivity could be compared with that of the new lavas. Measurements on all the samples were made within a few weeks after their collection and again in 1956.

The rocks were crushed to pass through a $\frac{1}{8}$ in. sieve and 120 cc of the material was filled into an annulus formed by a thin-walled (30mg/cm^2) β-sensitive geiger tube $1\cdot5$ cm diameter and a copper tube $3\cdot5$ cm diameter. The geiger tube was 15 cm long and when placed inside a lead castle gave a background of 21 pulses per minute. The standard deviation of the β count rate was kept less than 5%.

To check the potassium contribution to the β count rate, two standards, each containing 1% K_2O, were prepared by adding analytical potassium permanganate and potassium sulphate respectively to an inert material. These in turn were filled into the annular space around the geiger tube.

The rock samples, after β counting, were spread evenly over a shallow copper tray having an area of 25 sq. in., and placed inside a rectangular ionization chamber. Disintegration from α particles, leaving the surface of the sample, produced pulses of ionization which were recorded graphically. Only pulses were counted that resulted from the complete collection of the ionization created by the absorption of an α energy equal to or greater than 1 Mev.

Provided the tray was completely covered with a sample, it was observed that differences in particle size did not give rise to any significant variation in counting rate. The results obtained were reproducible within the standard deviation of the count rates, which was kept below 10%.

The ionization chamber was calibrated with two crushed rock samples of known Ra content supplied by the U.S. National Bureau of Standards. The samples used were Colombia River basalt, and a gabbro-diorite, containing respectively $0\cdot33 \pm 0\cdot03 \times 10^{-12}$ and $0\cdot18 \pm 0\cdot02 \times 10^{-12}$ g of Ra per gram or rock.

On White Island a survey was made with a geiger tube, 18 in. long and 1 in. diameter. The tube was always laid horizontally on the material to be examined,care being taken to place the counter where it would not be shielded from cosmic rays by the crater walls.

RESULTS

The standard potassium samples produced β count rates of $23\cdot0$ and $22\cdot8$ per min. respectively.

The standard radium samples gave α count rates of $3\cdot0$ and $1\cdot5$ counts per min. for the Colombia River basalt and the gabbro-diorite respectively, giving $9\cdot1$ and $8\cdot3$ counts per min./10^{-12} g of Ra/g of rock respectively.

The α and β count rates of the Ngauruhoe and Tongariro lavas are given in Table 1. For comparison are listed the results obtained from a sample of Haparangi rhyolite (Grange, 1937) and the β count rates given by Gibbs and McCallum (1955) for Taupo ash and Egmont ash which are of rhyolitic and andesitic composition respectively.

The results of the geiger tube survey of White Island are given in Table 2. The standard deviation of the count rate is of the order of 7%. For comparison with the andesites of White Island, typical pulse rates are given for Ngauruhoe and Tongariro lavas. All pulse rates have been corrected for a cosmic ray contribution of 120 pulses/min.

TABLE 1.—Radioactivity of Ngauruhoe Volcanics, Xenoliths Found in the 1954 Ngauruhoe Lava, Tongariro Lavas, and Haparangi Rhyolite, as Shown by α and β Count Rates.

Sample No.	Rock	α c/m	β c/m	K_2O*
1	Lava older than 1700 years† } from Ngauruhoe	1·0	28	—
2	Lava erupted in 1869	1·7	34	—
3	Lava erupted 12 Sept. 1954	1·3	35	1·1
4	Ngauruhoe lava 12 Sept. 1954. From this sample the vitrified gneissic xenolith sample No. 9 was separated	0·8	35	1·0
5	Ngauruhoe lava 12 Sept. 1954. From this sample the quartzose xenolith sample No. 10 was separated	2·6	34	1·1
6	Large boulder ejected from Ngauruhoe 1954	1·1	33	1·0
7	Ngauruhoe lava erupted 18 August 1954	2·5	31	1·0
8	Ash erupted from Ngauruhoe 1954	1·8	30	1·0
9	Vitrified gneissic xenolith separated from Ngauruhoe lava 1954, 12 Sept. The lava is represented by sample 4.	2·5	49	1·5
10	Quartzose xenolith separated from Ngauruhoe lava 1954, 12 Sept. The lava is represented by sample 5.	0·6	34	0·1
11	Lava older than 1700 years from Red Crater, Tongariro†.	1·1	20	—
12	Lava of unknown age from Red Crater, Tongariro	1·7	22	—
13	Haparangi rhyolite	2·9	87	—
14	Egmont ash, andesitic, Taranaki (C Horizon)‡	—	37	—
15	Taupo ash, rhyolitic, loamy sand, Ngakura, Rotorua (C Horizon)‡	—	79	—

*Determinations of K_2O by J. A. Ritchie, Dominion Laboratory.

†Lava was covered by Taupo ash.

‡According to Gibbs and McCallum, 1955.

TABLE 2.—Count Rates, Corrected for Cosmic Ray Contributions, From a Geiger Tube Laid Horizontally on Geological Formations on White Island (Samples 1 to 10), Ngāuruhoe (Sample 11), and Tongariro (Sample 12).

Sample No.	Description of Site	γ c/m
1	Lava at Troup Head S.E. of factory	57
2	Troup Head andesite breccia probably same lava as 1 on top of ridge leading to Troup Head.	55
3	Red andesite breccia resting against Troup Head on west side, on top of ridge.	50
4	Sample same as 3—a little lower	50
5	Light brown stratified tuff and ash on crest of Troup Head ridge overlying red breccia of 3 and 4.	44
6	Troup Head lava near highest point	57
7	Stratified tuffs underlying red breccia between factory and Wilson Bay.	38
8	Large andesite boulder in lahar mound. This lahar came from the west wall of the crater.	111
9	Boulder of andesite lava from south wall of crater.	75
10	Same as 9, but different boulder.	80
11	Ngauruhoe lava.	68
12	Tongariro lava.	63

DISCUSSION

A characteristic feature of the 1954 Ngauruhoe lava is the abundance of various xenoliths of gneissic origin (Steiner, 1958). Thus, sample 9 is intensely vitrified and contains relics of quartz and corroded, decomposed feldspar, whereas sample 10 consists predominantly of quartz. Both these xenoliths were comparatively large, and were separated from the enclosing lava for testing purposes. In addition, the lava contains numerous minute recrystallized xenoliths, essentially feldspathic in composition; they are also of gneissic origin. The feldspathic xenoliths are commonly only of microscopic dimensions. Because of their minute size, it has not been practicable to remove the feldspathic xenoliths, prior to testing, from the enclosing lava.

It has been shown earlier (Steiner, 1958) that the chemical composition of the vitrified gneissic xenoliths approaches that of the rhyolites of the Taupo volcanic association; that the rhyolite magma was derived by re-melting of acid gneiss; and that a reciprocal interaction took place between the magma and the feldspathic xenoliths, the magma acquiring a limited amount of silica, alumina, alkalies, and lime from the xenoliths by selective assimilation. It may be, therefore, that a limited amount of radioactive elements migrated in the opposite direction, i.e. from the magma into the re-melted xenoliths, thus depleting the magma in radioactive material.

From Table 1 it is evident that the values for the α count rates of the lavas and ash erupted from Ngauruhoe in 1954 vary considerably, ranging from 0·8 to 2·6 c/m. This variation seems to be essentially due to the presence of a varying amount of small feldspathic xenoliths or their absence respectively. The low α count rate of sample 4, i.e. the lava from which the intensely vitrified gneissic xenolith (Sample 9) was separated, seems to suggest depletion of the host magma, in radioactive elements with their consequent concentration in the gneissic xenolith (Sample 9). This would agree with the postulated limited migration of radioactive elements from the host magma into the xenoliths.

The quartzose xenolith, sample 10, has low α and β count rates that agree with other measurements made on quartz (G. J. McCallum, pers. comm.). Further, it should be noted that the vitrified gneissic xenolith, sample 9, and the Haparangi rhyolite, sample 13, show high α count rates, i.e. 2·5 and 2·9 c/m respectively. These high values appear to be in agreement with the hypothesis explaining the origin of the rhyolitic magma by re-melting of acid gneiss.

From the K_2O calibration and K_2O analyses (Table 1) it appears that the β count rates are mainly due to potassium. Count rates, given in Table 1, show that the ratio of α/β count rates has a limited range. These results are in agreement with the evidence obtained by Adams (1954) who showed that high values of radioactivity tend to be associated with high potash content in the lavas of the Lassen Peak region, California.

With the exception of the boulders, the γ count rates of the White Island rocks are generally lower than those for the rocks from Ngauruhoe and Tongariro. The lava flows forming the southerly and westerly crater walls (boulder samples 8, 9, and 10) are considerably more recent (J. Healy, pers. comm.) than the tuffs and breccias of the Troup Head. Except for the more recent lava flows, there is little difference between the radioactivities of the various White Island rocks. The recent flows show a distinctly higher level of radioactivity than the older breccias, tuffs, and lavas of Troup Head. The surface tuffs, consisting of comparatively fine material, show slightly less radioactivity than the breccias; this maybe attributed to leaching.

CONCLUSIONS

Conclusions derived from the data obtained are:

(1) The variation in α count rates of the lavas and ash erupted from Ngauruhoe in 1954 is due to the presence or absence respectively, of a varying amount of small feldspathic xenoliths.

(2) The radioactivity of the Ngauruhoe lavas is roughly proportional to the potash content.

(3) The recent lavas on the White Island are more radioactive than the older lavas.

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A SURVEY OF THE TEKTITE PROBLEM

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Summary

Present knowledge of the occurrence, properties, and composition of tektites is reviewed; and the principal theories which have been advanced to explain their origin are critically discussed. Nininger's lunar impactite theory, modified in accordance with the suggestions of Varsavsky is most free from objection, but if Kopal's contention that tektites must have had an origin closer to the Earth is well founded, Hardcastle's plastic sweeping theory seems likely to repay closer attention than it has so far received.

Introduction

Small lens- or button-shaped pebbles of silica-glass have been found in many widely separated localities where the geology of the region would make their occurrence unlikely. F. E. Suess (1900) named these bodies *tektites* (fused or molten stones), and suggested that they were of meteoritic origin. Writers on meteoritics generally accept this view; but some geologists and geochemists favour a terrestrial origin. The problem is complicated by the occurrence of volcanic glass, fulgurites, and meteoritic impactites, as well as of a number of objects of an artificial nature, all of which have been reported as tektites. When these "false tektites" have been eliminated from consideration, however, there still remains a large quantity of material which, if it is not of extra-terrestrial origin, appears at least to be the result of the action of some extra-terrestrial agency.

The composition and age of meteoritic material are significant data for a wide range of geophysical and cosmological studies. Since the conclusions that can be drawn depend greatly upon the supposed origin of the material, and upon the relative abundance of meteorites of different chemical composition, the problem of the true nature of the tektites is one of considerable interest. Meteorites are commonly supposed to have formed part of a disrupted planet, or to be samples of the raw material from which planets have been formed by accretion. Tektites are often considered to be an exception to this general principle. This paper is an attempt to weigh the merits of some currently accepted views of their origin and cosmological importance.

Occurrence of Tektites

Tektites have been reported from many countries, and they are usually assigned a local name according to their place of origin. These localities

N.Z. J. Geol. Geophys. 2: 183-94.

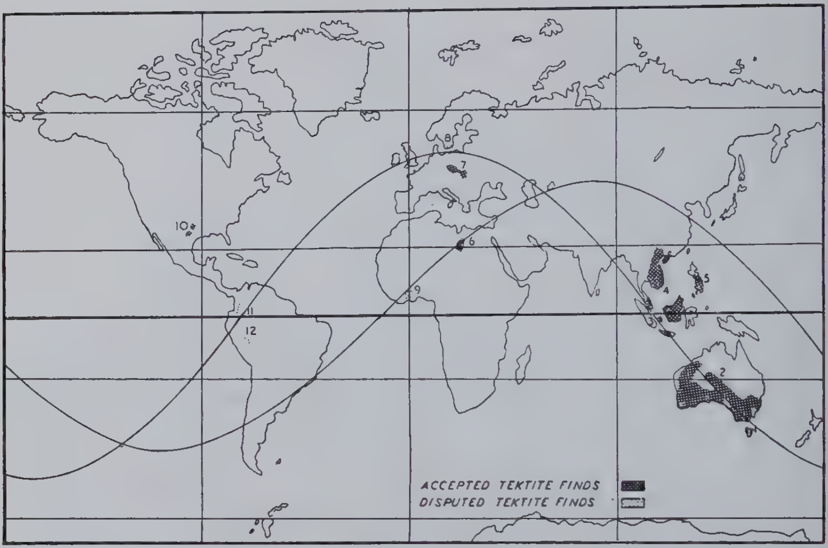

FIG. 1.—Distribution of principal tektite fields, showing their relationship to the great circles of La Paz.

(1) Darwin Glass
(2) Australites
(3) Billitonites
(4) Indochinites
(5) Philippinites, Rizalites
(6) Libyan Desert Glass
(7) Moldavites
(8) Schonite
(9) Ivory Coast Tektites
(10) Bediasites
(11) Colombianite
(12) Peruvianite

are identified in Fig. 1. The first tektites to attract attention were the moldavites, which occur in the south of Czechoslovakia, and were at first regarded as the remnants of a primitive glass factory. By the early years of this century, similar finds had been reported from Australia (australites), and Malaya (billitonites). At this time, several of the theories of origin still current were first advanced. Other places from which they have been reported include Indo-China (indochinites), the Philippines (philippinites), Texas (bediasites), Tasmania (Darwin glass or queenstownites), the Ivory Coast of Africa, and the Libyan Desert. Among the more doubtful cases are the americanites of Colombia and Peru, which are now considered to be obsidian (Friedman *et al.*, 1958), and the Swedish schonite, which Suess has identified as an industrial product (Lacroix, 1932, p. 10). The literature is very extensive. Barnes (1940) gives a bibliography containing nearly 250 references, and usefully summarizes the main facts about each find, as well as reproducing the locality maps from the original papers. In an important series of papers on australites, Fenner (1934, 1935, 1938, 1940, 1949b) also discusses the wide range of artificial objects which have been reported as tektites. Lacroix (1932) contains a particularly fine series of collotype plates.

The geological setting of the different tektite finds is very varied. Moldavites are found in Tertiary or Pleistocene quartz gravels, overlain in some places by silicified fresh-water limestones, and in others by a variety of soils (Suess, 1900). Australites are found exposed on the central plains of Australia, and also in the alluvial mining areas of Victoria and New South Wales. Billitonites are found in tin-pits, Ivory Coast tektites in gold-bearing gravels, and Darwin glass in the Tasmanian mining field on Mt. Darwin. The concentration of tektites in alluvial deposits is very usual, but the derivation of the gravels and the rocks on which they rest are both of very diverse character. Lacroix (1932) mentions granite, dacite, basalt, schist, gneiss, marble, and a variety of sedimentary rocks as old as Silurian. In view of the superficial resemblance of tektites to obsidian, most writers comment on the absence of volcanism. On the other hand, this resemblance may well mean that true tektites in volcanic regions have not been correctly identified.

Early proponents of a meteoritic origin stressed that the known deposits lay close to a single great circle, and presumably fell at one and the same time. The suggestion was even made that this great circle was the equator at the time of the fall. However, the discovery of the bediasites and the Ivory Coast tektites has made this view untenable, although an attempt was made by La Paz (1938) to revive it by postulating two falls, and hence a second great circle. Nininger (1952), who advocates a lunar origin, describes the falls as limited to within 40° of the equator, which appears to be rather too narrow a limit. Reference to Fig. 1, which shows both the tektite finds and the great circles of La Paz, will show that attempts to set geometrical limits to tektite distribution are somewhat forced, as it is quite possible for extensive falls in high latitudes, in the oceans, in volcanic areas, and in the Southern Hemisphere to have so far escaped detection. Since the falls themselves show no simple geometrical pattern, and since there are systematic physical and chemical differences between the various families of tektites, it seems beyond reasonable doubt that the occurrences are separate events, and it is not possible to draw valid conclusions about the nature of the orbits in space from the distribution of the presumed falls alone.

Composition and Physical Properties

Tektite material is generally described as a silica-glass. The silica content ranges from 70% in billitonites to $97\frac{1}{2}\%$ in Libyan Desert glass. Other constituents in excess of 1% are the oxides of aluminium, iron, magnesium, sodium, calcium, and potassium. The specific gravity ranges from 2·2 to 2·5, and the refractive index (for the sodium line) from 1·46 to 1·52. In colour they are generally a dark green or black, but lighter greens, amber, and brown are occasionally found, and an isolated indochinite is a yellowish-white. Details of many analyses have been listed and discussed by Barnes (1940), and are usefully summarized in tabular form by Stair (1955).

When tektites are cut, they are found to be markedly inhomogenous, containing internal bubbles and striae; and examination by polarized light reveals a strained condition. The flow structures and the flanged and button-like shapes are generally taken to indicate that they have travelled through the atmosphere at high velocity whilst in a molten condition. This appearance led many workers to class the tektites with igneous rocks, but Barnes (1940) pointed out that their composition was more nearly duplicated by that of certain argillaceous sedimentary rocks (Rankama and Sahama, 1942, p. 222). This is probably the most important argument advanced by those who favour a terrestrial origin. Stair (1955) points out, however, that the temperature of formation must have been between 1500° C and 2000° C, which is too great to be attributed to any volcanic processes.

Meteorites are generally classified into two main groups, according to their chemical composition. The *siderites*, or irons, are composed predominantly of iron, with the addition of some 10% of nickel, and rather less than 1% of cobalt. The *aerolites*, or stones, contain some 20% of iron, but their principal constituent is silica, which forms about 40%. A large sub-class of stones, the *chondrites*, are so named because of the presence of numerous granular inclusions, or chondrules, which are commonly of a glassy nature. In some chondrites, glass also forms an important constituent of the matrix. Meteorites of intermediate composition (sometimes called *siderolites*) also exist; but it is clear that tektites lie beyond this classification rather than within it. Nininger (1952), in reviewing the various analyses which have been made, rightly stresses the difficulty in arriving at an "average" composition for the various classes of meteorite.

Silverman (1952) has studied the ratio of the isotopes O^{18}/O^{16} in stony meteorites and tektites. Two stones (both achondrites) gave ratios of 6·3‰* and 6·5‰, whilst the tektites (a moldavite and an "americanite" from the Philippines) both gave 10·4‰. This ratio is within the range given by sedimentary rocks. The isotopic composition of their lead content also falls within the sedimentary range (Friedman *et al.*, 1958).

THE AGE OF TEKTITES

No geologist has suggested an origin for the tektites farther back than the Miocene (2×10^7 years), and nearly all have been assigned to Pleistocene or Recent by those geologists who have examined them *in situ*. Janoschek (*see* Barnes, 1940, p. 483) describes the moldavites as Helvetian (middle Miocene) in age, and the australites are undoubtedly Recent. Barnes (1940) relates the bediasites to the Jackson (Eocene) group of sedimentary beds. Studies based upon their argon content (H. E. Suess, 1952) indicate an upper limit of between 10^7 and 10^8 years. From their hard glossy composition, tektites might be supposed able to withstand terrestrial weathering for much longer periods than this, and it consequently appears justifiable to assume that no tektite is older than 10^8 years.

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*‰ = parts per thousand.

Both the geological evidence and the systematic differences in chemical composition between the different falls make it unlikely that tektites were produced by an isolated event which could not be repeated. The question therefore arises whether tektites still fall from time to time. The fact that no fall has been observed is inconclusive, since it cannot be reasonably suggested that falls are frequent events. It is also relevant to draw attention to the big discrepancy between the proportions of stony and iron meteriotes deduced from "finds" and "falls". Whereas some 90% of the finds are siderites, almost the same percentage of the observed falls are aerolites. It is consequently unlikely that evidence of the age of a particular fall (as distinct from a maximum age) will shed much light on their origin. On the other hand, if a fall were proved to have taken place within historic times, certain otherwise plausible theories could be ruled out.

THE ORIGIN OF TEKTITES

Theories of Terrestrial Origin

With the theories of terrestrial origin must be grouped those theories which invoke a collision between the Earth and a comet or meteorite, but which derive the material of the tektite from terrestrial rocks. The argument for terrestrial origin based upon composition is a strong one. Tektites have a chemical and isotopic composition which is very close to that of sedimentary rocks, and it seems unlikely that the surface of any other planet could have experienced a chemical cycle sufficiently similar to that involved in terrestrial sedimentation (Urey, 1958). Even if it were admitted that a rare igneous rock might approach the composition of a tektite, terrestrial igneous rocks are in a large part re-fused sediments, and the original difficulty is not removed. The real limit to his argument is set by the unsatisfactory state of present knowledge of the chemical behaviour of silicate melts.

There is no doubt that tektites have been subjected to considerable heat, and that volcanic sources are unable to produce a sufficiently high emperature to account for their formation. Barnes (1940) has argued for lightning as a possible source of heat; but an exhaustive study of objects known to fulgurites (Fenner, 1949a) shows them to be markedly different in physical form. When the wide range of geological settings n which tektites occur is considered, they would be likely to exhibit much wider range of composition and appearance if they were indeed f fulguritic origin.

Spencer (1933) considers that tektites are terrestrial impactites, resulting from the fall of a meteorite, apparently a siderite. The transformation of kinetic energy into heat is supposed to have raised the urrounding rock to a temperature of the order of 3500° C, and to ave exploded and sprayed molten gobbets to a considerable distance rom the point of impact. Under these conditions, tektites could be expected to show a marked relationship in composition to that of the rock on which they fall. Desert sands, on account of their richness in silica, are considered to be particularly favoured localities; and Spencer considers that tektites formed in this way should contain nickel from the meteorite as an impurity. Reported tektite analyses do not refer to nickel, "presumably because it was not suspected, and not looked for". Since none of the analyses quoted leaves more than ½% of unexplained material, it seems unlikely that there could be enough nickel present to support Spencer's hypothesis.

The wide area of distribution of the tektites is a more serious objection to terrestrial impactite theories. The australites, for example, cover an area about 4,000 km in length, and 2,000 to 3,000 km in width, comprising in all about two-thirds of the entire continent of Australia. The moldavites are less widely distributed, but they cover an area some 300 km by 200 km. Although transport by water can be adduced as a factor in distributing the moldavites, it seems probable that they have been concentrated rather than dispersed by the process. In the case of the australites, this mechanism cannot be invoked. The explosion of a single meteorite, even if it were very large, could not distribute the tektites so widely, and it would be necessary to postulate a shower. Since much of the area is dry, and the fall is believed to be Recent, explosion craters could be expected in appreciable numbers. Nininger (1952, p. 304) gives the maximum distance to which material is known to have been ejected from a terrestrial crater as 16 miles (25 km), and an upper limit would in any case be set by the velocity of escape. If the heat of impact is supposed to have completely destroyed the meteorite, it would be expected to have influenced the composition of the resultant impactites. The distribution of the tektites is at some variance with that suggested by Whipple and Rhinehart (*see* Varsavsky, 1958; *and* Friedman *et al.*, 1958).

Friedman *et al.* (1958) point out that only in the case of the Libyan Desert glass has the true size and shape of the strewn field been mapped, and suggest that a possible relationship with a group of supposed impact craters some hundred kilometres to the south should be studied. The distribution of the australites has, in fact, been well studied by Fenner and others, but the southern limit is set by the coastline. It will be of the greatest interest to see whether the intensive bottom sampling investigations now undertaken for oceanographic purposes will reveal new tektite fields, or a seaward extension of the known ones.

Disrupted Planet Theory

Stair (1955) considers that both the tektites and the other classes of meteorite had their origin in a disrupted planet. The planet originally occupied an orbit between Mars and Jupiter, and the cause of disruption was a collision between it and a moon of Jupiter. This cosmic disaster is considered to have given rise not only to the meteorites, but also to the asteroids and to comets. The tektite material is assumed to have been derived from a "glassy shell" of the planet. The Earth has no similar shell, because of re-working by the process of erosion and sedimentation. There is no evidence that any planet has ever possessed a shell of this nature; but, if it did, its properties could be expected to resemble those of an igneous rock rather than that of a sediment. Processes akin to sedimentation are expressly excluded by the theory.

Cometary Impact Theory

The suggestion of Urey (1957) that tektites were formed in the course of an encounter between the Earth and a comet is the most recent theory of tektite formation. The spectra of comets indicate the presence of compounds which would be chemically explosive when concentrated to higher densities (Watson, 1956, p. 66; Lyttelton, 1953, p. 54). On entering the atmosphere with a velocity of the order of 40 km/sec compression and heating would occur, and at a height of 60 to 100 km the explosive reaction would take place. No deep penetration of the surface would be expected, but terrestrial material would first be fused over a wide area, and then scattered by the explosion. Blasts of high-temperature gas during the process seem to Urey a more reasonable mechanism for producting tektite forms and flow structure than rapid passage through the atmosphere.

The possible frequency of cometary impact is estimated as one in every 5×10^7 years. This is too infrequent to account for the number of separate tektite fields. Urey therefore adopts a suggestion by Treanor that cometary nuclei may break up on approaching the Earth, and the separate masses fall in widely scattered places. If it is assumed that all tektites except the bediasites resulted from a single collision during the Pleistocene, and that the bediasites are Eocene, a time interval of the correct order is obtained.

During the explosion, tektites of varying composition are supposed to form; but, owing to the rapid crystallization of basic materials, only those of acidic composition could be expected to have the glassy nature necessary for long survival. The rapid heating and cooling involved in this process would enable tektites to retain their content of alkali metals, which would otherwise be expected to volatilize. Urey appears to believe that there is silicate material present in the comet, scattered in the form of fine dust; but it is not clear how great a part this material is supposed to play in determining the composition of the tektites, or what evidence there is for such a cometary composition.

The theory appears to offer an explanation of many of the observed characteristics of tektites, but further examination of the mechanism of the explosion is desirable, and the assumptions about cometary structure and composition require further justification. Factual information about comets is, indeed, still very meagre. Urey follows the comet model of Whipple (Watson, 1956, pp. 69–72), assuming a head some 10 km in diameter and a density of 0·01 g/cc. A kinetic energy of 5×10^{28} ergs is derived by assuming the standard meteor velocity of 42 km/sec. The chemical energy is estimated at perhaps 10^{27} ergs, but this must be subject to considerable uncertainty. If the nucleus of a comet contains discrete particles of the same order of density as rocks, but fairly widely separated (Lyttelton, p. 47), the mass of explosive gas may be several orders less than that estimated by Urey.

A catastrophic event of this kind would be expected to leave conspicuous traces, even though it may be difficult to predict the exact form they would take. Both lunar and terrestrial meteorite craters are evidence that the greater part of the debris from the explosion would be symmetrically distributed; and it might be expected that the configuration of the tektite fields would give an unmistakable indication of the impact area. In order to produce the observed distribution of the australites, the impact could not have taken place in the limestone Nullarbor Plain, which contains no siliceous rock whatever (Fenner, 1934), and if an impact in the Central Desert is assumed, it would be expected that tektites would also have been found to the north, which is not the case. An impact in the sea does not appear a likely alternative, but it seems the only way in which to explain island occurrences like the philippinites and the billitonites. A scheme of fragmentation of the cometary nucleus which would produce falls of widely different configurations in both hemispheres seems to call for very special pleading. On the one hand, it is well established observationally (Lyttelton, pp. 41–42) that the nucleus contracts as a comet approaches the Sun, and on the other, it appears that instability leading to prior tidal disruption is also highly probable (Lyttelton, p. 49). Finally, tektites seem from their gas content to have been formed at low pressures (H. E. Suess, 1951). This could hardly be the case if they were terrestrial material fused by cometary impact and dispersed by explosion. Whilst such uncertainties exist, it is not possible to give unqualified approval to a cometary hypothesis.

It is perhaps worth considering whether the tektites could have been cometary constituents. Meteoric material is scattered around orbits which are either those of known comets, or are of cometary type. Meteor showers occur when the Earth intersects these orbits. However, it is becoming increasingly clear that a distinction must be made between meteors and meteorites. Meteorites do not appear to be related to shower meteors (Lyttelton, p. 59 *et seq.*), and are consequently unrelated to comets. There is evidence that meteorites have been formed under pressure conditions of the kind likely to be present in a body of planetary size (Lyttelton, p. 148). If comets are formed by the accretion of intergalactic dust and its gravitational capture by the sun, as Lyttelton seeks to prove, meteoric material will be very different in kind from meteorites, and quite unlikely to show affinities with sedimentary rocks.

The Lunar Impactite Theory

Nininger (1952) has suggested that tektites might be lunar impactites, a by-product of the formation of lunar craters in the manner suggested by Gifford (1924, 1930). The small regular craters, and those with radiating ray systems are amongst the last of the lunar features to have been formed ; and the absence of similar markings on the Earth is taken to indicate formation at an early stage of geological history. Most writers consider that the earliest impacts took place whilst the Moon was in a molten or plastic condition, and that the maria are the result of a flow of material from a molten interior. Urey (1951) considers that meteoritic impacts could have raised the already solid lunar surface to some 2000° C, forming large lava pools which he identifies with the maria. This process would imply that lunar lavas would attain a higher temperature than any resulting from terrestrial volcanism. This material probably became solid between 2 and 4×10^9 years ago, which is, in geological terms, well into the pre-Cambrian. Subsequent falls upon these fused surfaces would produce explosion pits, and at least some of the material ejected from the crater could be expected to attain velocities in excess of the low velocity of escape (2·4 km/sec) characteristic of the Moon. This material is assumed to have reached the Earth as tektite falls, and tektites are therefore to be regarded as samples of the lunar surface. O'Keefe (1958) and Varsavsky (1958) have computed orbits which could bring this material to the Earth; but, whilst O'Keefe's proposals involve a slow spiralling taking some 10^7 years for the journey, Varsavsky considers that a direct transfer could take place. The shorter time-scale appears the more probable, as showers of tektites in the form of small localized clouds between the Earth and the Moon would be rapidly dispersed.

Gold (1958) has considered another aspect of the lunar theory. If the high temperatures needed for the formation of tektites are to be communicated to any considerable mass of material during the brief duration of an impact, the heat transfer cannot take place by conduction, and only a few millimetres of the outer surface would be melted. Gold points out that, provided the substance possesses suitable elastic properties, the energy could be rapidly distributed by a pressure pulse. This removes the further objection that, in order to produce a small quantity of liquid by conduction, considerable matter would have to be volatilized. Sand or powder derived from hard rocks has the necessary elastic properties.

A lunar origin of tektites would account for their form, their distribution, and their gas content, which other extra-terrestrial theories do not. Nevertheless, objections can be raised. Barnes (1958) points out that O'Keefe's slow spiralling to earth could be expected not only to produce continuing falls, but to have distributed the tektites throughout more of the geological column than is, in fact, the case. To Varsavsky's theory, which does not require so long a time-scale, he objects that it is unreasonable to expect nearly pure silica of the type of Libyan Desert glass to occur on the surface of the waterless, airless Moon. Urey (1958) points out that igneous processes which could produce very acid materials like the tektites would be very different from those experienced on Earth. However, he is forced to admit that the chemistry of silica is still very imperfectly known.

An even more serious objection is that of Kopal (1958), who points out that the ejecta from an impact explosion would have a wide range of velocities and angles of ejection. The passage to the Earth would result in further dispersal, and the material would be unlikely to land in the restricted fields in which the tektites apparently do. In his opinion, the origin of the tektites must be sought much closer to the Earth than the Moon, and Urey's cometary hypothesis seems to him most promising.

Kopal proposes an interesting experimental test. If the tektites are lunar impactites, they may be expected to show similarities to the rays from lunar craters. Recent Russian work has shown that the rays from Aristarchus are luminescent under certain conditions of solar radiation. If tektites are found to exhibit similar luminescence when bombarded with 50 to 100 keV protons, it would offer strong support to the lunar impactite theory.

The rays from lunar craters offer a further support to the impactite theory. If they consist of material ejected from the crater, then they show from their long straight course that at least a proportion of the material was concentrated in a very restricted range of azimuth. The circular form of the craters themselves is evidence of uniform ejection of material at the lower velocities, but it is possible that liquid droplets or blasts of gas at velocities above the velocity of escape might not conform to the statistical average of behaviour. This could invalidate Kopal's objection.

Plastic Sweeping Theory

Hardcastle (1926) has published an interesting suggestion which does not seem to have received detailed consideration, and of the writers quoted in the present paper, only Fenner seems to be aware of it. Hardcastle believes that tektites have been formed from stony meteorites by the process of ablation during their passage through the atmosphere. This process is not at all well understood. During its flight, a large meteorite loses from a third to one-half of its mass in a process which may include vaporization, melting, flaking, or mechanical destruction. Extreme pressures and temperatures are rapidly developed, and, although the efficiency of the transfer is probably low, it may be assumed that the rate at which mass is lost will be proportional to the rate at which work is done in decelerating the meteorite. Meteorites which have been observed to fall are generally accompanied by a dense persistent trail consisting of ablated debris (Rhinehart, 1957).

The surface of a stony meteorite is usually found to be pitted with holes of the same general order of size as that of tektites, and the whole surface of the stone is glazed. There are often traces of flow-structure indicating the predominant orientation of the stone during flight. It has not been possible to find any reference to an analysis of this glazed surface material alone, but it is reasonable to suppose that it would contain a selective sample of the material of the meteorite, rather than reflect its average composition. There is still considerable uncertainty about the chemistry of silica melts, and an analysis of this outer glazed material might be expected to show whether chemical differentiation of the right kind could take place. It is not necessary to suppose that the ablated material is all removed in molten form, because the data on the relative frequency of "finds" and "falls" indicate that the stony material is both impermanent and difficult to recognize. By postulating either single falls or meteorite showers, the configuration of the various tektite finds can be readily explained. Systematic differences in the average composition of the tektites in different regions is to be expected, since they are derived from different parent meteorites.

The possibility of deriving tektites from other types of meteorite has been considered by Suess and Lacroix (*see* Fenner, 1938, p. 209). They suggest that tektites are the residual silica from a meteorite composed predominantly of the light metals, such as aluminium, which would be completely burned out during their passage through the atmosphere. However, evidence for the existence of bodies of this kind is lacking; and no tektites have been found with inclusions of the light metals.

If Hardcastle's theory were accepted, it would imply that tektite falls had not ceased; but they would necessarily be rare events, since either very large stones or a concentrated shower would be needed to produce the quantity of glassy material required. Tektites are not completely immune to weathering, as a study of the moldavites and billitonites will clearly show; but they are sufficiently durable to make the fact that the earliest tektites appear to be Miocene and objection to all theories which allow continuing falls.

CONCLUSIONS

No completely satisfying account of the tektites has yet been given. Of the theories which derive them from terrestrial material, only Urey's cometary theory appears to provide an agent sufficiently powerful to account for formation and distribution. Unfortunately, the present gaps in knowledge of cometary structure are so large that they present an obstacle to more than qualified acceptance. Of the extra-terrestrial hypotheses, Varsavsky's form of the Lunar impactite theory seems open to the fewest objections. Some of the objections which have been raised to the impactite theory are based upon assumptions about the nature of the Moon's surface, and may therefore be dismissed as begging the question. However, Kopal's objections based upon the difficulty of maintaining a stable cloud of meteorites in space have more substance. Hardcastle's theory goes some distance to meeting this point; and even if it does not prove adequate to explain the tektites, the attention it draws to the mechanism of ablation appears to open up an interesting field for further study.

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SOME FOSSILS FROM HITHERTO UNDIFFERENTIATED PERMIAN-TRIASSIC-JURASSIC ROCKS NEAR AUCKLAND

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(Received for publication, 14 October 1958)

Summary

Some fossils are recorded from hitherto undifferentiated Permian-Triassic-Jurassic greywacke sandstones and argillites near Auckland. *Inoceramus* cf. *galoi* and *Buchia ?malayamaorica* from the Hunua Ranges and *I.* cf. *galoi* from Ponui Island indicate an Upper Jurassic age for these rocks; a pelecypod resembling a Permian *Atomodesma* or a Jurassic *Inoceramus* is recorded from Kawau Island.

INTRODUCTION

During field trips conducted by the Auckland University Field Club, fossils were discovered by members at three localities near Auckland (Fig. 1) in areas shown as undifferentiated Permian-Triassic-Jurassic sediments in maps published by the New Zealand Geological Survey in 1949 (1:1,013,670) and 1958 (1:2,000,000).

FOSSIL LOCALITIES

In April 1958 Mr J. G. Colebrook, Auckland University, found a fossiliferous pebble in alluvium immediately below the junction of the Mangatangi and Konini streams in the Hunua Ranges, 25 miles south-east of Auckland city (N.Z. fossil record No. N48P/598, grid ref. 708340). The two fossils in the pebble were identified by Mr I. G. Speden of the New Zealand Geological Survey as *Buchia ?malaya-maorica* Krumbeck and *Inoceramus* cf. *galoi* Boehm. Although the fossils were not found *in situ*, the catchment basin above the river fork covers approximately only 20 square miles and it is almost certain that the parent rock is upstream. The sediments within this basin can therefore be regarded provisionally as Upper Jurassic in age.

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In August 1956 the writer found four casts of *Inoceramus* cf. *galoi* Boehm exposed on the shore platform on the east coast of Ponui Island, some 20 miles east of Auckland city (N43P/501, grid ref. 707614; N43P/502, grid ref. 708616). This suggests a Heterian age for these rocks. The strike of the beds in this section of the coast varies from 135° to 155°; dips are to the south-west and range from 28° to 65°. The rocks are bedded or massive medium to coarse greywacke sandstones, microbreccias, and argillites.

In April 1955, a distorted pelecypod cast was found by the writer on the south-east coast of Kawau Island (N34P/503, grid ref. 388108). Dr C. A. Fleming (*in* Hopgood 1956) stated that the poor preservation prevented discrimination between Permian *Maitaia* (now regarded as a

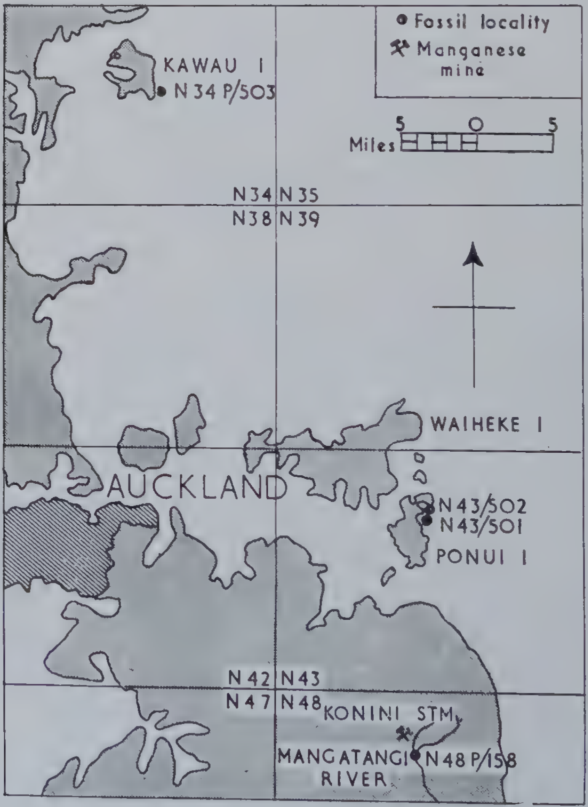

FIG. 1.—Map of the Upper Jurassic and ?Permian fossil localities near Auckland.

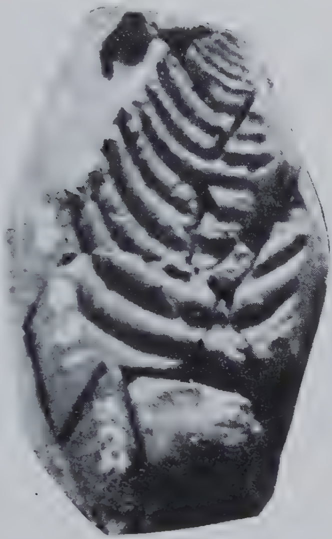

FIG. 2.—*Inoceramus* cf. *galoi* Boehm. Ponui Island. N43/502. Coll. E N. Milligan 1956. Dimensions of impression: 70 mm × 45 mm.

synonym of *Atomodesma* Waterhouse, 1958), and Jurassic *Inoceramus*. Despite the uncertainty in age the occurrence indicates the possibility of finding better preserved material in these apparently unpromising rocks.

ACKNOWLEDGEMENTS

Thanks are expressed to Mr J. G. Colebrook for permission to record his discovery, and to Mr S. B. Seagar for supplying details of structure in the Ponui area where he is now carrying out research.

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THE WAREPAN STAGE (TRIASSIC): DEFINITION AND CORRELATION

By J. D. Campbell, University of Canterbury

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Summary

The Warepan Stage is defined and the sequence at the type locality (Kaihiku Stream, South Otago) is described. Criteria for correlating beds in other sequences with those at the type locality are critically examined. Fossils described from Warepan horizons are listed.

Introduction

The Warepan Stage of Marwick (1951) was based on the type section of Ongley's (1933) Warepa Series for which Allan (1945, p. 5) had noted "further field work is obviously necessary". In his description of the type area, Ongley (1939, p. 41) recorded "only one fossiliferous outcrop . . . in the Warepa Series"; he considered the Otapiri Series to be absent and defined both upper and lower boundaries of the Warepa Series on lithological criteria.

Mr G. Warren and the writer spent a week in the region about Lochindorb Station in December 1954; within a restricted area it was found possible to add some refinements to Ongley's mapping.

Historical

Warepa Series was proposed by Ongley (1933, p. 15) for that part of the old unit Wairoa Series containing *Monotis richmondiana* Zittel.

In 1878 Hector proposed Oreti Series to replace the Lower Wairoa Series as used by Cox (1878) and McKay (1878a) in their reports on the Hokonui Hills, and from this time onwards until 1932, Wairoa Series was used by most geologists to cover the group of beds ranging from those with *Mytilus problematicus* Zittel up to and including those with *Monotis richmondiana* (McKay, 1879, p. 101; Hector, 1882: index of fossil localities; Hutton, 1885, p. 204; Hector, 1886, p. 72 and caption of fig. 41; Hector, 1892: index of fossil localities; Park, 1910: captions of pls. II and III). Beds with *Mytilus* but not those with *Monotis* were included in the series by Trechmann (1918, table facing p. 237), Henderson and Ongley (1923, p. 23) and Wilckens (1927, p. 44), while Benson (1921, p. 24) included only *Monotis* beds in the series.

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In 1932 Williamson proposed a two-fold division of the Wairoa Series as follows: Whakahau Series for beds with *Mytilus problematicus* and Marakopa Series for those with *Monotis richmondiana*. The new units were based on sections exposed about 3 miles south of Marakopa River on the South Auckland coast. The Marakopa Series, conformable on Whakahau beds, was stated to be succeeded by beds correlated with the Otapiri Series.

Ongley's subdivision of the Wairoa Series into Otamita Series (beds with *Mytilus problematicus*) and Warepa Series was proposed a year later. The series were based on Southland and South Otago sections but were closely similar in content to those of Williamson.

Marwick (1946, p. 27-9) retained Williamson's classification of rocks in the Te Kuiti Subdivision. He noted the equivalence of Williamson's and Ongley's series and stated "if series names are applied to the whole Dominion, then Williamson's having a year's priority, should be used; but if adjectival stage names are introduced they should be based on the stratigraphically clearest and most fossiliferous localities and not limited by priority."

In setting up stages, Marwick (1951) followed his own injunction and rejected the older names. He did not justify rejection of Marakopa in terms of the requirements mentioned in 1946, indeed it would be difficult to do so in light of the then known stratigraphy. Completeness of exposure in a coastal section could be a factor favouring "clear stratigraphy" and with the key fossil occurring profusely "throughout several hundred feet of strata", Marakopa was obviously the "most fossiliferous" locality. A reason for rejecting Marakopa in favour of Warepa might have been nearness to type sections of other Triassic stages, all of which are located in Southland and Otago.

Marwick (1953) fixed the type locality for the stage and listed its fauna and distribution.

The material from which the key fossil, *Monotis richmondiana* Zittel 1864, was described was collected by Hochstetter in 1859 from a "hill slope, east of Richmond", Nelson.

Trechmann (1918) noted and figured *M. richmondiana* as well as nine varieties, of which three were new, of *Monotis* and *Pseudomonotis*, from Warepan localities. (As a result of Marwick's (1935) and Muller's (1938) work all would now be placed in *Monotis*.)

Marwick (1935) reviewed Trechmann's systematics and included all his forms in *M. richmondiana*. In 1946 the same author recognized three distinct forms of *Monotis* at Marakopa in a bed considered to be Otapirian. One of these he described in 1953 as *M. calvata* Marwick.

DEFINITION

The Warepan Stage is here defined as those beds laid down at the type locality after the appearance of *Monotis richmondiana* Zittel and before the appearance of an Otapirian fauna (including *Spiriferina*

(*Rastelligera*) *diomedea* Trechmann). In stratal terms the stage consists of approximately 1,200 ft of beds between localities S179/582 and S179/585.

In the type section the following are restricted to Warepan beds:

Spiriferina (*Rastelligera*) sp.
Clavigera sp.
Palaeoneilo cf. *mundeni* Fleming
Monotis richmondiana Zittel

The following are rstricted to beds that are Warepan by correlation (see below for localities):

Terebratula spp. (2)
Coenothyris sp.
Spiriferina aff. *trechmanni* (Wilckens)
S. (*Psioidea*) cf. *nelsonensis* Trechmann
Athyris sp.
Myophoria sp.
Monotis calvata Marwick
?*Torastarte* sp.
?*Pleurotomaria* sp.

Type Locality

After examining the type sections of Williamson's Marakopa Series and Ongley's Warepa Series the writer favours the latter as a basis for the time-rock unit between Otamitan and Otapirian Stages. The *Monotis*-bearing beds at Marakopa are involved in a complicated system of folds and thrusts and in spite of the completeness of exposure the structural complexity of the coastal section makes it quite unsuitable as a type locality.

The type section of the Warepan Stage (Marwick, 1953, p. 17) follows a NE/SW line along a tributary* of Kaihiku Stream about a mile NW of Rocky Dome, Warepa Survey District, Sheet S179, South Otago. It consists of outcrops in strike ridges on the watershed between the Kaihiku and Puerua drainage systems.

The main ridge of Rocky Dome and its north-eastern slopes are composed of Otamitan rocks; *Mytilus problematicus* occurs in outcrops along the lower course of the tributary mentioned above (S179/569-574). The *Mytilus*-bearing siltstones are succeeded by *Halobia*-bearing siltstones with brachiopods (*Athyris* cf. *wreyi* (Suess) and *Mentzeliopsis* sp. at S179/578) and these by a massive sandstone with abundant plant remains which forms the Rocky Dome ridge. *Halobia* sp. and *Palaeocardita quadrata* Trechmann were collected from an horizon (S179/577) high in the massive sandstone.

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*The tributary can readily be identified from the N as it is the first stream NW of Rocky Dome to breach the ridge.

The bed in which *Monotis richmondiana* makes its first appearance in the section immediately overlies the ridge-forming sandstone; it is preserved in medium grained blue-grey sandstone with green "mud pellets" and pyrite nodules. Five outcrops of the basal bed were seen in the tributary, all west of its notch in the ridge. In each outcrop there is a profusion of shells of *Monotis* mostly preserved as separated valves and with much fragmentary material.

The greatest thickness of shell bed exposed in any one outcrop is 6 ft (S179/582). Undescribed species of *Rastelligera* and *Clavigera* were the only other forms collected in the basal bed but only a few blocks were available for breaking up.

A pocket of *M. richmondiana* was discovered in a grey fine sandstone (S179/583) 30 ft stratigraphically above the basal bed. Following a break in the section there is an outcrop of blue-grey siltstone 170 yards S of S179/582; the horizon is about 390 ft above the basal bed and *Palaeoneilo* cf. *mundeni* Fleming and crinoid stems were collected (S179/584).

Basal Otapirian forms were collected from an outcrop (S179/585) about 9 chains S of S179/582 in fine-grained blue sandstone approximately 1,200 ft stratigraphically above the base of the Warepan. The outcrop occurs in a prominent strike ridge N of the bush line and the following were collected:

Nuculana sp.
Myophoria sp.
Mytilus sp.

The species of *Myophoria* listed has been noted from basal Otapirian collections elsewhere (Campbell, 1956, p. 48). An Otapirian age for the faunule is further strengthened by species represented in collections from about the same horizon as S179/585 at outcrops between 2 and 3 miles SE along the strike. The following were collected beside a hillside track N of Lochindorb homestead (S179/587, 588, 589):

S. (*Rastelligera*) sp.
Myophoria sp.
Triaphorus sp.
Pteria cf. *contorta* (Portlock)
Mytilus sp.
Anodontophora sp.

S. (*Rastelligera*) *diomedea* and *Clavigera tumida* were collected from S179/592, 40 chains SE of Puerua River, in an horizon thought to be equivalent to S179/585.

CORRELATION

In keeping with views expressed by the writer elsewhere (Campbell, 1956, p. 45), the Warepan Stage is *defined* in terms of the incoming of a single species (*Monotis richmondiana*). If, in a number of widely separated sections, individual beds marked by the incoming of this species are to be *correlated*, then some special justification is called for. In both islands and at distances from the type section of up to 800 miles the writer and others have mapped units termed "Warepan Stage" of which the lower boundary is placed at the incoming of *Monotis richmondiana*. The lowest bed in each of the units termed "Warepan Stage" can be time-equivalent only if *M. richmondiana* colonized New Zealand seas instantaneously, geologically speaking. If this were so, two main conditions must have been satisfied: the organisms must have possessed a highly developed means of dispersal and, secondly, they must have been tolerant of a wide range of bottom conditions. The dramatic appearance of *M. richmondiana* in section after section in New Zealand and New Caledonia and of closely related species around the Pacific Ocean and beyond suggests that at least the first condition was met (Allan, 1956, p. 368). The species is preserved in abundance in siltstone, sandstone, conglomerate, and limestone, representing a wide range of bottom conditions. Correlation on the *first appearance* of *M. richmondiana* would seem, then, to be justified.

At least locally, the *disappearance* of *M. richmondiana* seems to have been haphazard. In Taringatura Hills, on a northern spur of Taringatura Hill (S159/718) *Monotis* shell beds are succeeded by a cream vitric tuff and this by further beds with *Monotis*. Rather less than 2 miles N, a similar tuff overlies *Monotis* shell beds (S159/721) but *Monotis* does not appear above it in the section.

From time to time *M. richmondiana* has been used zonally (Wellman, 1952, p. 24; Marwick, 1953, p. 18), its presence alone being taken to indicate deposition of the bed concerned during Warepan time. While this is a sound procedure in areas that are structurally simple, it is hazardous in areas where the relationship of the *Monotis* beds to other fossiliferous horizons is uncertain. It must be noted that Marwick (1946, p. 30) mapped as Otapirian, beds with *Monotis* spp., fragments of which cannot be distinguished from *richmondiana*.

The Warepan Stage is not merely the *richmondiana* zone. The disappearance of the key fossil in the type sequence plays no part in the definition of the stage, and there, as in many other sequences, beds high in the Warepan do not contain the key fossil. Nevertheless, the presence of *M. richmondiana* in abundance in Upper Triassic sequences must remain the basis *par excellence* for correlating units of rock as Warepan in age.

There are several Upper Triassic sequences in which no Warepan rocks have been recognized, Otapirian beds succeeding Otamitan (E. Hokonui Hills, Watters, 1952, p. 474; Nugget Point, Campbell, 1955, pl. 39; Gore, Wood, 1956, p. 62). At Nugget Point the writer has observed a break in exposure of 60 ft immediately N of and below the basal Otapirian shell bed. Neither there nor in any other section have undoubted Otapirian rocks been shown to lie directly on undoubted Otamitan beds. On a broad regional basis, however, published work on Otago and Southland Triassic stratigraphy suggests that *M. richmondiana* beds—but not necessarily Warepan beds—are lensoid. Within

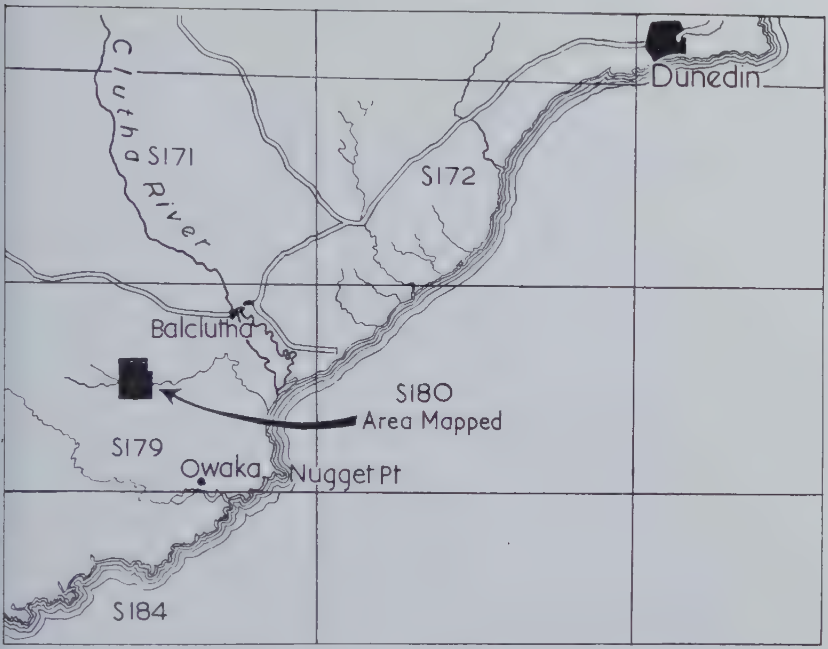

FIG. 1.—Locality map.

the area included in the map above (Fig. 1) the key fossil for the stage has been found only in the NW corner. In all sections SE of S179/582 there is a break in exposure between Otapirian and Otamitan beds.

THE WAREPAN FAUNA

The explanation of the absence of *Monotis* beds from some Upper Triassic sections may be that there, a disconfromity represents their position as suggested by Watters (1952, p. 474) and Wood (1956, p. 62). It is clear that near-contemporary deposition and erosion of *Monotis* beds occurred at Marakopa where an unconformity (Marwick, 1946, p. 28) separates two sets of these beds; the upper set contains much fragmentary shell material and, in the writer's opinion, could have been derived from that below. *M. calvata*, described from above the unconformity, occurs in profusion below it and although time is undoubtedly represented by the unconformity the writer has found no positive evidence favouring an Otapirian age for the *Monotis* beds above the unconformity.

M. richmondiana is widely distributed in the region about Arthurs Pass and is almost invariably preserved in rocks which by virtue of their poor sorting as to size, are likely to be the products of submarine slumping. The abundance of *Monotis* shells in some beds such as those at S66/572, SW of Bealey, suggests derivation from shell beds like those of the type section.

Very detailed mapping is needed in the Southland sections without *Monotis* beds before the presence of a disconformity can be established but a pre-requisite to this work must be an appreciation of the complete Warepan fauna known so far.

Notwithstanding McKay's statement to the contrary (1878B, p. 126) the *Monotis* shell beds in Southland, Nelson, and Awakino-Marakopa-Kawhia contain a varied fauna, some members of which, like *M. richmondiana* and *M. calvata*, are not known to occur in Otamitan or Otapirian rocks. Further palaeontological work will undoubtedly lead to the recognition of a distinct Warepan fauna.

In the Taringatura Hills, in a shell bed in which *Clavigera* is conspicuous in the outcrop (S159/637) the following occur:

Terebratula spp. (2)
Spiriferina aff. *trechmanni* (Wilckens)
S. (*Psioidea*) aff. *nelsonensis* Trechmann
S. (*Rastelligera*) sp.
Clavigera sp.
Myophoria sp.
?*Myophoria* sp.
Monotis richmondiana Zittel
?*Pleurotomaria* sp.
orthocone

Some of the same forms have been recorded from Warepan localities in the north Taringatura Hills by Coombs (1950, p. 438).

Coenothyris sp., figured by Trechmann (1918), occurs in *Monotis* shell beds in many Nelson localities; *S.* (*Rastelligera*) sp. is abundant in *Monotis* beds at G.S. 432, Ram Creek. At Hart Creek (S20/579) the following occur:

Terebratula sp.
Coenothyris sp.
Spiriferina aff. *trechmanni* (Wilckens)
S. (*Psioidea*) aff. *nelsonensis* Trechmann
S. (*Rastelligera*) sp.
Clavigera sp.
Monotis richmondiana Zittel

At Marakopa, 6 ft below the unconformity in *Monotis* beds (N82/526), *M. richmondiana* and *M. calvata* are accompanied by *Athyris* sp. and ?*Torastarte* sp. Immediately above the unconformity (N82/528) the same forms occur along with

Spiriferina sp.
Clavigera sp.
Triaphorus sp.

In the Awakino area Mr J. A. Grant-Mackie (pers. comm.) has collected *Rhynchonella* sp. from *Monotis* shell beds (N91/558).

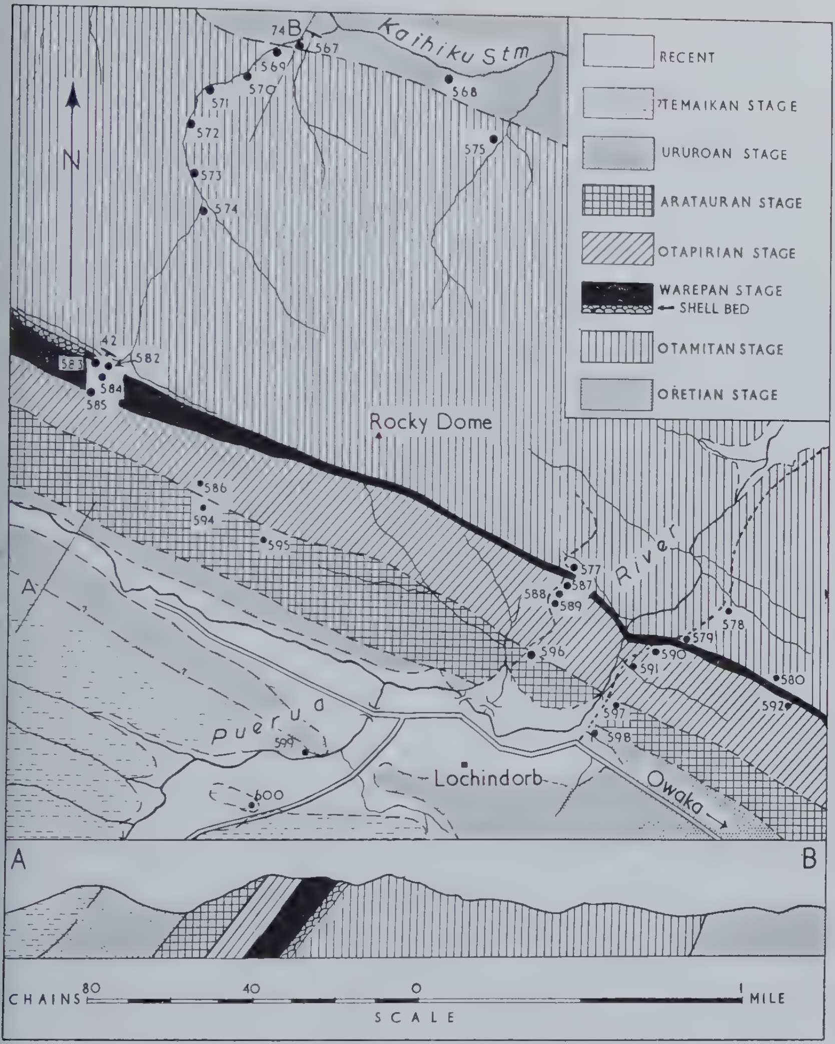

FIG. 2.—Map and Section of part of Warepa Survey District, Sheet S179, South Otago. Fossil localities are indicated by fossil record form numbers for S179, e.g., 582.

WAREPAN FACIES

The ratio of shell material—much of it finely comminuted—to sand, silt, and mud in the *Monotis* shell beds of Southland, Otago, Nelson, and Auckland is commonly high enough to form very calcareous clastics but true limestones are not developed. In Canterbury, however, *Monotis* shell limestone, a truly organic sediment, occurs at Okuku (Campbell and Warren, 1955). *Monotis* also occurs in tuffaceous limestones at Okuku along with a variety of other forms (polyzoans, brachiopods, pelecypods) of which enough is known to say that they are not met with in less calcareous rocks. It is here that the concept of two distinct facies within the New Zealand Triassic might be developed with a true biological basis*.

Essentially contemporaneous, contrasting faunal assemblages but each with *Monotis richmondiana*, flourished in distinct environments, comparatively lime-free on the one hand (Otago, Southland, Nelson, South Auckland: the Hokonui Facies of Wellman, 1952), lime-rich on the other (Okuku: part of the Alpine Facies). Complication of this simple picture is suggested by the restriction of *M. calvata* to the North Island and Canterbury and the apparent absence in the North Island of forms of the basal Otapirian assemblage of Otago and Southland (Campbell, 1956) although some do occur in Nelson (Ram Creek, G.S. 197, S20, 518, 584).

Faunal Index

The following species have been described from beds that are Warepan by correlation:

Monotis richmondiana K. Zittel, 1864
M. calvata J. Marwick, 1953

Acknowledgements

Thanks are expressed to Mr G. Warren for his help in the field and for criticism of the MS, and to Mr and Mrs W. J. McNab, Lochindorb Station, for their hospitality.

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*The writer put forward this concept in a paper entitled "Calcareous facies of the older rocks of North Canterbury" presented at the 8th N.Z. Science Congress, 1954. It should be noted that records of *Monotis*, based on fragments of ribbed shell, at Kaiwara, Gore Bay, and Waiau Gorge are incorrect and must be deleted.

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NEW ZEALAND ^{14}C AGE MEASUREMENTS—4

By G. J. FERGUSSON and T. A. RAFTER, Division of Nuclear Sciences, Department of Scientific and Industrial Research, Lower Hutt

(Received for publication, 11 December 1958)

This list gives details of samples dated in the Nuclear Sciences Laboratory, in the period from June 1956 to July 1958, and follows on from previously published results (Fergusson and Rafter, 1953, 1955, 1957).

The method that has been used is proportional counting of carbon dioxide gas, as described by Fergusson (1955) and Rafter (1955a). Operational procedure and computation of errors, etc., is virtually unchanged from that given with the previous list.

The same modern wood standard as used previously has also been used for this series of measurements but consideration is being given to a better standard of contemporary ^{14}C activity. Due to the Suess effect (Suess, 1955) the ^{14}C specific activity of the present New Zealand wood standard has been reduced by 1·50% so that 120 years has been added to ages of samples that lived on the land, calculated according to this modern standard. For samples other than wood or charcoal the appropriate correction for fractionation effects (Rafter, 1955b) has also been made.

The ^{14}C specific activity of modern surface ocean carbon has also been reduced by the Suess effect, but so far it has not been possible to determine accurately the magnitude of the reduction. However, consideration of data by Craig (1957), Brannon *et al.* (1957), and Fergusson (1958) on the Suess effect dilution of surface ocean carbon indicates that the most likely value was 1·25% at the time when the New Zealand wood standard was cut. A correction of 100 years has therefore been added to all ages of surface ocean material calculated according to the New Zealand modern wood standard.

ACKNOWLEDGEMENTS

The authors wish to acknowledge the co-operation of Mr H. Polach and Miss J. Gower who have carried out the preparation of the carbon dioxide and also of Messrs H. James, K. Gibbons, and R. Penman who have operated the counting equipment. The assistance of Mr T. Grant-Taylor with geological aspects of the samples is also gratefully acknowledged.

N.Z. J. Geol. Geophys. 2 : 208-41.

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N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
79	R 50	Ministry of Works, Auckland, N.Z.	N42/283605	N42/530
80	AA 2276/1	M. T. Te Punga, N.Z. Geological Survey, D.S.I.R.	N149/033427	N149/523
81	AA 2276/2	M. T. Te Punga, N.Z. Geological Survey, D.S.I.R.	N148/796206	N148/503
82	AB 1188/2	C. J. Banwell, Dominion Physical Laboratory, D.S.I.R., N.Z.	N94/533460	N94/519
83	AB 1188/3	V. J. Chapman, Auckland University College, Auckland, N.Z.	—	N42/531
84	R 20	B. W. Collins, N.Z. Geological Survey, D.S.I.R.	S84/995558	S84/504
85	R 21	H. W. Wellman, Victoria University College, Wellington, N.Z.	N164/336229	N164/542
86	R 91/2	B. W. Collins, N.Z. Geological Survey, D.S.I.R.	S84/009519	S84/521
87	AB 1455	R. W. Willett and G. J. Lensen, N.Z. Geological Survey, D.S.I.R.	N164/336214	N164/555
88	R 36	I. C. McKellar, N.Z. Geological Survey, D.S.I.R.	—	S115/500
89	T 2525/2	P. Wardle, University of Otago, Dunedin, N.Z.	S164/100880	S164/504

Locality	Sample Details	Age. Years before 1950
Civic Square, Auckland, N.Z.	Carbonized wood from well-defined carbonaceous seam 1-3 in. in thickness, approx. 25 ft below present ground level. 10 ft below upper level of Waitemata beds.	> 43,000
Dairy Factory well, Awahuri, N.Z.	Wood from depth of 233 ft in well.	> 45,000
Well near water tower, Foxton, N.Z.	Wood from depth of 155 ft in well. Sample thus 150 ft below sea-level. (*See* Te Punga, 1958: *N.Z. J. Geol. Geophys. 1* : 92-4.)	9,900 ± 150
Cave at west end of Alum Lake at head of Waiora Valley, Wairakei, N.Z.	Wood from log 4 in. diameter and 2 ft in length.	2,000 ± 100
Between West Tamaki Head and Brown's Island, Auckland, N.Z.	Peat sample from 6 ft bed at 110 ft depth in a submarine bore.	> 37,000
Public Hospital, Christchurch, N.Z.	Wood from boiler foundation excavation depth 11-15 ft. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.)	1,550 ± 80
Terrace—Bowen Street, Wellington, N.Z.	Wood from depth of 65 ft in drill hole for building foundation.	> 37,000
Bowenvale Road, Christchurch, N.Z.	Wood, 8 ft below road level, south abutment of Heathcote Bridge. (*See* Suggate, 1958. *N.Z. J. Geol. Geophys. 1* : 103-22.)	940 ± 70
Herbert Street—20 ft from west corner Dixon Street, Wellington, N.Z.	Wood from depth of 35 ft in drill hole for foundations of building. In probable Upper Pleistocene terrace.	> 40,000
Lake Hawea, N.Z.	Peat—bottom bed of peat deposited against old slump—Lake Hawea outlet. Date will be slightly younger than commencement of retreat of N.Z. glaciers at the end of the last glaciation.	15,100 ± 200
Silver Peaks District, Dunedin, N.Z.	Wood — *Podocarpus totara* from 2,300 ft altitude on hillside now carrying snow tussock and silver beech.	650 ± 60

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
90	T 2525/9	G. T. S. Bayliss, University of Otago, Dunedin, N.Z.	S155/120930	S155/502
91	T 920/3	C. A. Fleming, N.Z. Geological Survey, D.S.I.R.	N137/239967	N137/552
92	T 920/4	C. A. Fleming, N.Z. Geological Survey, D.S.I.R.	N138/553837	N138/513
93	R 100/3	M. Gage, Canterbury University College, Christchurch, N.Z.	S74/338832	S74/523
94	R 100/4	M. Gage, Canterbury University College, Christchurch, N.Z.	S74/394868	S74/525
95 96	R 214/1 R 214/2	B. W. Collins, M. Gage, R. P. Suggate, and R. W. Willett, N.Z. Geological Survey, D.S.I.R.	S74/394868	S74/526 S74/527
97 98	R 214/3 R 214/4	B. W. Collins, M. Gage, R. P. Suggate, and R. W. Willett, N.Z. Geological Survey, D.S.I.R.	S74/338832	S74/528 S74/529
99	R 90/5	R. R. Stewart, Christchurch City Council, N.Z.	S84/972594	S84/516
100 101	R 83/1 R 83/2	H. W. Wellman, N.Z. Geological Survey, D.S.I.R.	N105/532295	N105/521

Locality	Sample Details	Age. Years before 1950
The Gap—Silver Peaks, Dunedin, N.Z.	Wood — *Podocarpus totara* from 1,850 ft altitude on hillside now carrying snowgrass, some silver beech in vicinity, but no other forest species.	670 ± 60
Mouth of Waitotara River, Waverley, N.Z.	Wood from drowned forest, tree trunks standing upright in river bed. Suggests a subsidence of about 10 ft in the last 1,000 years.	1,020 ± 60
Landguard Bluff, Wanganui, N.Z.	Wood from youngest band in lignite bed, Kaiwhara Alluvium, youngest member of the Rapanui Formation, deposited during a period of aggradation and slight climatic amelioration after a period of river entrenchment through the marine Waipuna Delta conglomerate.	> 45,000
South bank of Rubicon River, N.Z.	Peat, overlain by Kowhai River terrace. Sample will date the Kowhai River terrace.	6,050 ± 110
North bank of Joyce Stream, Canterbury Plains, N.Z.	Wood and peat from 200 ft below surface of Canterbury Plains. Sample will date the Otarama surface equivalent of Canterbury Plains.	> 45,000
North bank of Joyce Stream, ¼ mile upstream from ford on disused Otarama Road, N.Z.	About 200 ft below Otarama surface Fossil wood Peat	 > 42,000 > 40,000
South bank of Rubicon River, ¼ mile downstream from homestead at end of Rubicon Road, N.Z.	10 ft below surface of main Kowhai River terrace, i.e., on top of Otarama surface. Wood Peat	 6,050 ± 80 10,200 ± 120
Blighs Road, Christchurch, N.Z.	Peat. 150 ft depth in well. Sample to determine position of a hypothetical unconformity between post-glacial alluvial and marine sediments. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.)	> 45,000
North Shore of Lake Waikaremoana, 2¼ miles W.N.W. of Lake House.	Wood from trees drowned by rise of lake. Standing trunks were exposed by a lowering of lake level by about 50 ft in Jan. 1956. Samples date rise of lake level to modern position. Wood from one tree Wood from another tree	 2,200 ± 60 2,190 ± 60

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
102	R 160/1	J. W. Brodie, N.Z. Oceanographic Instiute, D.S.I.R.	N160/385348	N160/531
103	R 160/5			N160/532
104	R 160/4			N160/535
105	R 160/3			N160/536
106	R 160/2			N160/537
107	R 160/8	J. W. Brodie, N.Z. Oceanographic Instiute, D.S.I.R.	N160/446468	N160/534
108	R 160/6	J. W. Brodie, N.Z. Oceanographic Instiute, D.S.I.R.	N160/397395	N160/533
109	R 160/7A	J. W. Brodie, N.Z. Oceanographic Instiute, D.S.I.R.	N164/305224	N164/551
110	R 160/7B			
111	R 178	M. T. Te Punga, N.Z. Geological Survey, D.S.I.R.	N164/491272	N164/554
112	R 177	M. T. Te Punga, N.Z. Geological Survey, D.S.I.R.	N164/389251	N164/553
113	R 188/6	H. M. Pantin, N.Z. Oceanographic Institute, D.S.I.R.	39° 16′ S 177° 22′E	Ocean Station B4

Locality	Sample Details	Age. Years before 1950
Takapu Road, Johnsonville, N.Z.	Wood samples from clay fill in a V gully cut in greywacke. These samples will show rate and continuity of deposition of fill, and check the former morphology and drainage pattern of this area.	
	Wood from upper horizon of fill	21,500 ± 350
	Wood, duplicate of R160/1	21,300 ± 350
	Wood 2½ ft below R160/1	22,300 ± 350
	Wood 7 ft below R160/1	22,800 ± 350
	From angular conglomerate at base of fill. 14 ft below R160/1	37,500 ± 1,600
	(*See* Brodie, 1957 : *N.Z. J. Sci. Tech. B38* : 623-43.)	
Kahao Creek, N.Z.	Wood from layer of logs 3 ft below surface of aggradation terrace at present stream level. (*See* Brodie, 1957 : *N.Z. J. Sci. Tech. B38* : 623-43.)	1,650 ± 60
Porirua-Johnsonville Road, Porirua, N.Z.	Peat (N.Z. 74 > 45,000 yr) overlain by gravel deposit followed by a layer of clay with plant remains, a further gravel layer, clay with plant remains (this sample) and angular fragments above. (*See* Brodie, 1957 : *N.Z. J. Sci. Tech. B38* : 623-43.)	> 41,000
Karori, N.Z.	A suspected fossil soil 5 ft below the present surface overlain by weathered clay. Is probably of first interstadial age.	
	Washed sample	24,000 ± 500
	Sorted sample	27,000 ± 600
West bank of main drain, Fitzherbert Road, Wainuiomata, N.Z.	Wood with bark, size 8 x 4 x 3½ in. 5 ft of silt has been laid down above the wood since swamp formation ended. Will date close of swamp formation.	7,230 ± 100
One mile N.W. of main town bridge, Wainuiomata, N.Z.	Wood from branch 4 in. in diameter with bark intact. From peat section 5 ft thick. Sample 4 in. above greywacke bedrock. Will date commencement of swamp formation which postdates the last local episode of solifluction.	11,500 ± 160
Sea-bed of Hawke Bay, N.Z.	Rolled fragments of carbonized wood recovered from flattened pumice gravel —depth 30 fathoms. The age will give a maximum age for the environment of deposition, probably a former beach line.	2,030 ± 100

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
114	R 151/2	R. M. Cassie, N.Z Oceanographic Institute, D.S.I.R.	—	—
115	R 212	P. G. Bamford, Ohawe Beach, Hawera, N.Z.	N129/813299	N129/526
116	R 229/1	E. T. Annear, N.Z. Geological Survey, D.S.I.R	S44/900909	S44/828
117	R 229/2	B. W. Collins, N.Z. Geological Survey, D.S.I.R	S76/033773	S76/505
118	R 229/3	R. P. Suggate, N.Z. Geological Survey, D.S.I.R	S83/891322	S83/501
119	R 229/4	R. P. Suggate, N.Z. Geological Survey, D.S.I.R	S84/010565	S84/522
120 121	R 229/5 R 229/6	L. E. Oborn, N.Z. Geological Survey, D.S.I.R	S84/974563	S84/523 S84/524
122	R 229/7	Samson, Drainage Board, Christchurch, N.Z.	S84/040567	S84/525

Locality	Sample Details	Age. Years before 1950
Muriwai Beach, Auckland, N.Z.	Toheroa shells from sub-fossil deposit behind fore-dunes, 10 miles north of beach road. These shells are larger than contemporary shells and possibly belong to an extinct race.	1,030 ± 60
On property of D. Ward, Paora Road, Hawera, N.Z.	Wood in peat layer at depth of 35 ft in well. From last lahar deposits of Mt. Egmont.	> 40,000
Six ft above south side of main Greymouth-Reefton Road. Two hains towards Grey-nouth from Kamaka Station, North West-and, N.Z.	Wood from thin bedded sands and silts which are over and underlain by gravels. Material apparently dammed by aggradation from the Arnold Valley during the K2 advance of the Kumara sub-stage.	22,300 ± 350
Ioore's gravel pit, Kaiapoi, N.Z.	Shell. 30 ft depth in gravel pit. 15 ft below mean sea-level. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.)	6,800 ± 90
Dept. of Agriculture lot, Lake Ellesmere, .Z.	Wood. 75 ft depth in well. 71 ft below sea-level. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.)	9,400 ± 120
orner Madras and hester Sts, Christ-hurch, N.Z.	Wood. 73 ft depth in bore 57·5 ft below sea-level. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.	8,000 ± 150
larence Road, hristchurch, N.Z.	Samples from well drilled in Riccarton Borough. Shells, strongly weathered, 43 ft depth in well. 11 ft below sea-level. Peat 137 ft depth in well 105 ft below sea-level. (*See* Suggate, 1958 : *N.Z. J. Geol. Geophys. 1* : 103-22.)	6,200 ± 120 > 43,000
orner of Woodham nd Worcester Sts, hristchurch, N.Z.	Wood. 14 ft depth. 3 ft below sea-level. (*See* Suggate, 1958 :. *N.Z. J. Geol. Geophys. 1* : 103-22.)	3,810 ± 70

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
123	R 229/8	I. C. McKellar, N.Z. Geological Survey, D.S.I.R.	S169/425376	S169/710
124	R 90/3	F. E. Bowen, N.Z. Geological Survey, D.S.I.R.	S44/754967	S44/826
125	R 100/2	T. L. Grant-Taylor, N.Z. Geological Survey, D.S.I.R.	N115/577824	N115/520
126	R 199/4	T. L. Grant-Taylor, N.Z. Geological Survey, D.S.I.R.	N115/569835	N115/526
127	R 91/1	T. L. Grant-Taylor, N.Z. Geological Survey, D.S.I.R.	N134/263213	N134/600
128	R 199/1	G. J. Lensen, N.Z. Geological Survey, D.S.I.R.	N129/852240	N129/509
129	R 161/1	D. Kear and E. Pharo, N.Z. Geological Survey, D.S.I.R.	N58/645655	N58/502
130 131	R 81/1 R 81/2	N. E. Odell, University of Otago, N.Z.	S78/600560	S78/506
132	R 81/3	N. E. Odell, University of Otago, N.Z.	S78/330490	S78/507
133	R 79	R. E. Bell, University of Oklahoma, U.S.A.	—	—

Locality	Sample Details	Age. Years before 1950
Between Tanner and McKenzie roads, Hokonui, Southland, N.Z.	Wood, with bark present. From base of 5 ft thickness of peat in peat bog.	7,020 ± 100
Cliff overlooking Rapahoe Beach, N.Z.	Wood, soft and brittle with bark. 20 ft below top of gravel terrace, 100 ft above sea-level.	6,500 ± 100
Mohaka River, N.Z.	Wood, from basal 2 ft of last aggradation gravels. Will date commencement of earliest phase of aggradation in the last major aggradation period.	> 45,000
Mohaka River, N.Z.	Wood from 15 ft clayey horizon in Mohaka aggradation terrace, deposition of fine material following down-cutting in previously deposited gravels of last major fluvial aggradation period.	> 45,000
Hastings, N.Z.	Shells from shell-bed 97 ft depth in bore. From an intertidal estuary during an early phase of Flandrian transgression.	6,620 ± 100
Beach at eastern end of golfcourse at end of Fairfield Road, Hawera, N.Z.	Wood from peat layer immediately above top lahar.	1,450 ± 60
South-west coast of Mt. Maunganui, about 30 chains south of Tokatapu Point, N.Z.	Shells from shell-bed 10 ft above present sea-level. This beach has not been affected by pumice derived aggradation from Waikato River.	620 ± 80
Right bank Bullock Creek, 50 yds above road bridge approx. 5 miles south of Weheka, Westland, N.Z.	Wood, embedded in gravel 6 ft below surface.	Modern < 100
	Repeat run on another sample.	Modern < 100
Bruce Bay. One mile north of township, junction of new road and old shore road, South Westland, N.Z.	Wood, 3 ft below surface in 7 ft thick sandy marine terrace.	3,320 ± 120
Parawhakatau, one mile north of Claverley, Kaikoura Coast, N.Z.	Wood from wall post taken from house pit C. Legendary accounts suggest that this pa was occupied for only a short period (about 20 yr) and should date 250-300 years.	320 ± 60

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
134	R 80/1	R. E. Bell, University of Oklahoma, U.S.A.	—	—
135	R 80/2			
136	R 192/4	L. Lockerbie, Otago Museum, Dunedin, N.Z.	—	—
137	R 192/1			
138	R 192/8			
139	R 192/9			
140	R 192/3			
141	R 192/10	L. Lockerbie, Otago Museum, Dunedin, N.Z.	—	—
142	R 192/15			
143	R 192/6			
144	R 192/12			
143	R 192/14			
146	R 192/2	L. Lockerbie, Otago Museum, Dunedin, N.Z.	—	—
147	R 192/5	L. Lockerbie, Otago Museum, Dunedin, N.Z.	—	—
148	R 295/1	J. W. Brodie, N.Z. Oceanographic Institute, D.S.I.R.	N160/269305	—
149	R 295/2			

Locality	Sample Details	Age. Years before 1950
Mouth of Tahakopa River, Papatowai, Otago, N.Z.	Charcoal from the lowest levels of the occupational debris of a moa-hunter site.	765 ± 30
	Repeat run on another sample.	755 ± 30
Tahakopa River Mouth, Otago, N.Z.	Samples from an early Maori moa-hunter site. Surface cover now forest. Charcoal associated with moa and sea bones from bottom of trench.	630 ± 50
	Moa bone (*Eurapteryx gravis*) resting on black deposit of charcoal-ash-sand and associated with bones and actifacts. Approx. mid-way up the deposit.	460 ± 50
	Moa bone (*Dinornis maximus*) associated with seal bones and shells.	460 ± 80
	Moa bone (*Eurapteryx gravis*) associated with other moa and seal bones.	310 ± 60
	Moa bone.	390 ± 80
False Island, Otago, N.Z.	Refuse deposit lying on sand. Shells, fishbones, and charcoal associated with artifacts.	
	Shells, pipi	470 ± 60
	Fishbones	290 ± 80
	Charcoal (from oven)	320 ± 50
	Charcoal (not oven)	345 ± 50
	Charcoal (from oven)	215 ± 50
Tautuku, Otago, N.Z.	Moa bones—(*Dinornis torosus*) from Dry-top sandhill in occupation deposit. Moa bones lying immediately below human skeletons buried in clean top layer sand.	280 ± 80
Cannibal Bay, Otago, N.Z.	Shells (mussel) from deposit on sandhill slope, well above sea-level. Surface cover chiefly clean sand.	450 ± 60
West headland of Warehou Bay, Makara, N.Z.	From a Maori occupation site which had been abandoned before European times. No account preserved in Maori history.	
	Mollusc and Paua shells	Modern i.e. < 160 yrs.
	Wood from butt of palisade post	310 ± 60

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
150 151	R 188/1 R 188/2	J. W. Brodie, N.Z. Oceanographic Institute, D.S.I.R.	41° 50′ S 174° 26′ E	Ocean Station 445 Ej8
152 153 154 155	R 188/3 R 188/4 R 188/5 R 188/7	J. W. Brodie, N.Z. Oceanographic Institute, D.S.I.R.	41° 21′ S 174° 17′ E	Ocean Station CP 37
156	R 25	C. G. Vucetich and D. Cross, N.Z. Soil Bureau, D.S.I.R.	N66/308204	N66/501
157	AA 3384	J. Healy, N.Z. Geological Survey, D.S.I.R.	N94/695523	N94/506
158	R 18	T. L. Grant-Taylor, N.Z. Geological Survey, D.S.I.R.	N134/212181	N134/594
159	R 19	T. L. Grant-Taylor, N.Z. Geological Survey, D.S.I.R.	N134/206194	N134/593
160	R 26	R. J. Cameron, N.Z. Forest Service	N86/121816	N86/503

Locality	Sample Details	Age. Years before 1950
Sea-bed off Cape Campbell, Cook Strait, N.Z.	Semi-calcareous concretionary boulder dredged from depth of 70 fathoms. The boulder is highly fossiliferous, containing numerous shells embedded in a fine grained matrix consisting of lime and non-calcareous detritus. Fossil studies indicate that the shells were deposited during a period of cold climate.	
	Shells from boulder	19,500 ± 1,000
	Matrix of concretion	27,500± 3,000
Sea-bed—Cook Strait, N.Z.	Shells from sand and gravel samples around Cook Strait "narrows". It is probable that erosion of the sea-bed around the narrows has occurred within recent times so that shells ranging back as far as the last Glacial period may well be present on the sea-bed.	
	Shells (*Ostrea*)	Modern i.e. < 200
	Shells (*Venericardia*)	Modern i.e. < 250
	Shells (*Chlamys*)	580 ± 100
	Shells (Lamellibranchs) outside Cloudy Bay	250 ± 80
Putaruru-Rotorua Road, 3½ miles N.E. Putaruru, N.Z.	Carbon beneath Tirau ash shower.	33,800 ± 1,000
Taupo-Rotorua Road. ¼ mile N.E. of Kaimanawa Road, N.Z.	Charred wood 10-12 in. above Waimihia Lapilli member. 8-10 in. below base of Taupo Lapilli member.	2,270 ± 100
Pumice pit, Pakowai Road. ½ mile S.S.W. of Longlands Road, Hastings, N.Z.	Charcoal intimately mixed with Taupo pumice flood debris—hence date of eruption. Depth 1-10 ft below grass surface.	1,760 ± 80
From drain one chain east of unformed road between Irongate Road and Maraekakaho Road, Hastings, N.Z.	Wood from shrubs in growing position on clay. Plants killed by pumice flood, probably by rise in groundwater table—hence date of pumice flood.	1,750 ± 80
South side of Rawhiti Road—six miles N.N.W. of Fort Galatea, N.Z.	Wood from Totara log between Kaharoa and Taupo Pumice ash.	1,300 ± 80

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
161	R 31	J. Healy, N.Z. Geological Survey, D.S.I.R.	N84/371871	N84/501
162	R 33	J. Healy, N.Z. Geological Survey, D.S.I.R.	N84/371871	N84/502
163	R 34	J. Healy, N.Z. Geological Survey, D.S.I.R.	N85/533645	N85/501
164	R 35	J. Healy, N.Z. Geological Survey, D.S.I.R.	N85/533654	N85/502
165	R 37	J. Healy, N.Z. Geological Survey, D.S.I.R.	N75/134136	N75/502
166	R 197/6	D. Cross, N.Z. Soil Bureau, D.S.I.R.	N85/853841	N85/508
167	R 103/3	D. Cross, N.Z. Soil Bureau, D.S.I.R.	N85/783802	N85/507
168	R 197/1	C. G. Vucetich, N.Z. Soil Bureau, D.S.I.R.	N94/723463	N94/521
169	R 149/1	E. F. Lloyd, N.Z. Geological Survey, D.S.I.R.	N85/828834	N85/504

Locality	Sample Details	Age. Years before 1950
Left bank of sludge channel, Kinleith Mill —¼ mile from mill, N.Z.	Wood from topmost member of Taupo pumice.	1,780 ± 80
Sludge channel, Kinleith Mill, right bank, ¼ mile from mill, N.Z.	Charcoal from Taupo pumice. Small twigs and branches disseminated through the bed. Bed believed to be mud flow. Sample should date either mud flow or eruption before it.	1,830 ± 70
Pumice pit west side of Taupo-Atiamuri Road, N.Z.	Charcoal and twigs from dark grey pumice sand and rhyolite tuff and lapilli above. Taupo pumice. Probably dates eruption which is believed to have destroyed the vegetation.	1,840 ± 50
Pumice pit west side of Taupo-Atiamuri Road—one mile north of Maroa corner, N.Z.	Small twigs up to ½ in. diam. from uppermost bed in Taupo pumice deposits.	1,890 ± 70
Below spillway—Arapuni dam, Arapuni, N.Z.	Small branches, roots, twigs, and leaves. Vegetation on old soil at site of Arapuni buried forest.	1,900 ± 70
South bank of Lake Okaro, N.Z.	Wood. 20 ft depth in breccia. Remains of 70-year-old manuka tree.	840 ± 50
Pit two miles west of Waiotapu Hotel and near Lake Tutaeinanga, N.Z.	Wood, partially decomposed rimu log, embedded in altered andesitic ash erupted from site of Lake Ngapouri. Sample 48 in. below surface and 9 in. above top of Taupo ash bed.	3,110 ± 70
Road cutting on left bank of Pueto River at bridge on Broadlands, Taupo Road, N.Z.	Wood in dark brown peaty bed immediately below Hatepe Lapilli bed, 36 ft below soil surface. 3 ft above Pueto River surface.	1,900 ± 60
On western side of eroded gulch entering Lake Ngahewa crater. Rainbow Mountain, Waiotapu, N.Z.	Wood from outer surface of trees 2-3 ft in diameter lying directly on top of Taupo silty loams from (Taupo pumice shower—250 A.D.) and covered by 20 ft of hydrothermally altered acid andesite.	900 ± 40

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
170	R 199/7	D. R. Gregg, N.Z. Geological Survey, D.S.I.R.	N111/905769	N111/562
171	R 203/1	D Cross, N.Z. Soil Bureau, D.S.I.R.	N84/283703	N84/503
172	R 197/5	C. G. Vucetich and D. Cross, N.Z. Soil Bureau, D.S.I.R.	N86/958868	N86/506
173 174	R 203/2 R 203/3	J. Healy and B. N. Thomson, N.Z. Geological Survey, D.S.I.R.	N85/496736	N85/505
175	R 203/9	J. Healy, N.Z. Geological Survey, D.S.I.R.	N94/618489	N94/515
176 177	R 203/7 R 203/10	J. Healy, N.Z. Geological Survey, D.S.I.R.	N94/589347	N94/512 N94/513
178 179	R 203/5 R 203/6	J Healy, N.Z. Geological Survey, D.S.I.R.	N94/560439	N94/510 N94/511

Locality	Sample Details	Age. Years before 1950
National Park-Wanganui highway, one mile south of junction of National Park-Taupo Road, N.Z.	Charred wood collected about 4 in. from base of 14 in. cream rhyolitic pumice ash. This is overlaid by 10 in. Ngaurohoe Ash and underlaid by 50 in. Tongariro Ash and lapilli.	1,800 ± 50
Cutting on Whakamaru-Tihoi Road, N.Z.	Small branches, roots, twigs collected from top layer of an old well weathered ash shower. Should give a check on western sections of the Taupo ash beds, but this sample appears to date the "host" ash bed rather than the overlying ash bed, as intended.	2,650 ± 150
Brett's road near northern shore of Lake Rerewhakaitu, N.Z.	Carbonized root, embedded within 2 in. of the top of the Taupo Ash and therefore expected to date overlying ash. However, this sample appears to date the "host" ash bed, rather than the overlying ash bed as intended.	1,800 ± 100
Road cutting near Atiamuri State Hydro, N.Z.	Small branches and twigs collected from basal 6 ft of the coarse pumice debris member. Fragments have been carried down the Waikato River in a pumice lahar.	1,750 ± 50
	Sample from near base of underlying layer—6 ft fine pumice debris.	1,800 ± 100
Rotorua Road, 4 miles N.E. of Wairakei, N.Z.	Small charred branches and twigs from rhyolite block member. Sample will date the volcanic eruption and check whether this eruption was Taupo lapilli or an earlier one.	1,850 ± 100
Quarry on Napier Highway, 1¼ miles S.E. of Terraces Hotel, Taupo, N.Z.	Charcoal dust and fragments of organic material up to ¼ in. in size collected from top inch of ash member 9. Sample should date Hatepe Lapilli.	1,900 ± 70
	Charoal fragments up to ½ in. in size from top inch of ash member 11. Sample should date a Taupo eruption.	2,500 ± 200
Top of first rise on access road to Bore 203, Wairakei, N.Z.	Partially carbonized dust and fragments of organic material up to ½ in. in size coated with white sediment, collected from top ½ in. of ash member 9. Sample should date one of Taupo eruptions.	2,100 ± 100
	Charred twigs and branches ¼ to 1 in. long collected from top of ash member 19. Sample should date eruption Waimihia Lapilli.	3,420 ± 70

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
180	R 330/1	J. Healy, N.Z. Geological Survey, D.S.I.R.	N103/531143	N103/501
181	R 90/7	D. R. Gregg and E. F. Lloyd, N.Z. Geological Survey, D.S.I.R.	N112/150900	N112/520
182	R 203/4	J. Healy, N.Z. Geological Survey, D.S.I.R.	N94/695523	N94/509
183	R 103/1	H. S. Gibbs, N.Z. Soil Bureau, D.S.I.R.	N114/247843	N114/501
184	R 103/2	H. S. Gibbs, N.Z. Soil Bureau, D.S.I.R.	N114/247843	N114/505
185	R 197/3	C. G. Vucetich, N.Z. Soil Bureau, D.S.I.R.	N94/589347	N94/520
186	R 199/2	D. R. Gregg and H. Stewart-Killick, N.Z. Geological Survey, D.S.I.R.	N112/250808	N112/521
187	R 384	I. C. McKellar, N.Z. Geological Survey, D.S.I.R.	—	VL 1

Locality	Sample Details	Age. Years before 1950
Right bank of the Minemaiaia River upstream from Pakikokuru Creek, Taupo, N.Z.	Carbonized wood from charred log, collected 2 ft from base of the 60 ft pumice breccia member.	3,150 ± 90
Four chains from face of older Te Mari flow. Mangatetipua Stream in Okahukura bush, N.Z.	Wood, from ash layer beneath flow from Te Mari. Flow and sample overlie washed Taupo pumice.	Modern i.e. < 200
Rotorua-Taupo Road, ¼ mile N.E. of Kaimanawa Road corner, N.Z.	Charcoal up to ¼ in. in size with large amount of silt present, collected from near centre of 3-5 in. thick layer in member 9.	2,800 ± 100
Pohokura Road, 9 miles N.W. of Tutira, Hawke's Bay, N.Z.	Charcoal and coarse pumice sand from a layer 27-29 in. below the surface. Host layer is correlated with basal ash bed of the Taupo Shower, 50 miles distant.	1,800 ± 70
Pohokura Road, 9 miles N.W. of Tutira, Hawke's Bay, N.Z.	Charcoal and pumice sand from a layer 36-39 in. below the surface and immediately above ash bed classified as Waimihia Lapilli. Date indicate that this ash bed erupted at a much later period than Waimihia Lapilli and about the same time as the Tongariro Ash.	2,400 ± 80
Terraces Quarry—1½ miles S.E. of Terraces Hotel, Taupo, N.Z.	Carbon flecks and soil from a 5 in. thick bed immediately below Waimihia Lapilli bed.	8,850 ± 1,000
Cutting on Waiouru-Turangi Highway at bridge over Mangatawai Stream. One chain south of bridge on the south side of road, N.Z.	From lower 3 in. of 20 in. of leaf-bearing unweathered andesitic ash, which has possibly come from Ngauruhoe. Separated from overlying Taupo pumice by 13 in. of andesitic ash without leaves. This leaf bearing ash is a clearly defined marker bed over 150 square miles.	2,500 ± 200
Cape Hallett Base, Antarctica.	Young penguin remains 15 in. below surface on which penguins nest at present. The sample was lowest organic material in this section and rested on 6 in. of coarse sand and fine gravel. Sample should determine date at which this rookery was colonized. (*See* Harrington and McKellar, 1958 : *N.Z. J. Geol. Geophys.* *1* : 571-6.)	1,210 ± 70

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
188 189	R 154/1 R 154/3	J. M. L. de Azcona, Spanish Geological Institute, Madrid, Spain	—	—
190	R 153	M. A. Reynolds, Vulcanological Observatory, Rabaul, New Guinea	—	—
191	R 296/1	J. E. Coaldrake, C.S.I.R.O., Brisbane, Australia	—	—
192	R 296/2	J. E. Coaldrake, C.S.I.R.O., Brisbane, Australia	—	—
193 194	R 296/3 R 296//4	J. E. Coaldrake, C.S.I.R.O., Brisbane, Australia	—	—
195	R 296/5	R. M. Moore, C.S.I.R.O., Canberra, Australia	—	—
196	R 296/6	J. Penman, Irrigation Research Station, Merbein, Victoria, Australia	—	—
197	R 296/7	E. K. Duncan, Hunter Ditsrict Water Board, Newcastle, Australia	—	—

Locality	Sample Details	Age. Years before 1950
Leson Highway, Spain.	Graphite from bed, rich in graphite, interstratified with shales. Repeat run on another sample	> 32,000 > 37,000
Tunnel Hill Road, Blanche Bay, New Guinea.	Charcoal from lower part of pyroclastic deposits which form the Blanche Bay caldera wall. These deposits are about 50 ft thick where exposed and consist mainly of pumice dust and some bands of pumice lapilli at the base. Sample will date commencement of last catastrophic phase of activity.	1,190 ± 60
Beach front 14 miles south of Double Island point, Australia.	Wood *Podocarpus* embedded in "brown coal" at high water mark. A mangrove has been identified from same bed.	> 45,000
Beach front 4 miles south of Double Island point, Austraila.	Wood, myrtaceous not *Eucalyptus*, probably *Metrosideros*. From bed of compacted carbonaceous sand at high water mark. Evidence from this site suggests association with a former sea-level, together with a warmer climate.	> 40,000
Cliffs fronting beach 13 miles south of Double Island point, Australia.	Charcoal, embedded in a sandy matrix 250 ft above sea-level. Sample should establish a minimum age for Teewah sands. Organic cemented sand from the hardpan of a former ground water podzol. 10 ft above sample.	39,000 ± 3,000 30,300 ± 800
Eucania, near Babinda, Australia.	Mangrove peat, from flat a few feet above high water mark.	6,270 ± 120
Physche Bend, Mildura, Australia.	Wood. 35 ft depth in bore. 83 ft above sea-level.	5,400 ± 80
Newcastle, Australia.	Shell-mollusc *Anadara* from bore 29A. This sample will date the marine transgression and provide an upper limit for the time within which the dunes have been built.	> 33,000

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
198 199 200 201	R 264/1 R 264/2 R 264/3 R 264/4	B. E. Butler, C.S.I.R.O., Canberra, Australia	—	—
202 203	R 340/1 R 340/2	H. Field, Coconut Grove, Florida, U.S.A.	—	—
204	R 332	G. A. Taylor, Vulcanological Observatory, Rabaul, New Guinea	—	—
205	R 266/1	D. Mawson, University of Adelaide, Australia	—	—
206	R 288/2	D. Mawson, University of Adelaide, Australia	—	—
207	R 414/2	E. D. Gill, National Museum of Victoria, Melbourne, Australia	—	—
208	R 319	F. Engel, Abancay, 377, Lima, Peru	—	—
209 210	R 285 R 308	F. Engel, Abancay, 377, Lima, Peru	—	—
211 212	R 370/1 R 370/2	F. Engel, Abancay, 377, Lima, Peru	—	—

Locality	Sample Details	Age. Years before 1950
South-eastern Australia.	Soil samples containing fragments of wood and charcoal which were deposited with the soil and were not later intrusions.	
	K3 cycle	29,000 ± 800
	K2 cycle	3,740 ± 100
	K1 cycle	390 ± 60
	K0 cycle	Modern < 120
Baluchistan, Pakistan.	Charcoal from Kapoto rock shelter	Modern < 200
	Charcoal from mound. 28 miles south of Panjguar.	Modern < 200
Long Island crater, New Guinea.	Carbonized wood, from small tree trunk that was buried by a catastrophic eruption from Long Island crater.	Modern < 100
Lake Callabona, South Australia.	Crop contents of *Diprotodon*, fossilized in standing position evidently by being bogged in mud. The *Diprotodon* is a giant herbivorous marsupial which formerly lived in numbers in Australia, but appear to have died out suddenly at period of sudden climatic change.	> 40,000
Orroroo, South Australia.	Dentine from lower jaw teeth of *Diprotodon*.	6,700 ± 250
Keilor, 10 miles N.W. from Melbourne, Australia.	Charcoal with silt from aboriginal fireplace at the Keilor Cranium site—5 ft 9 in. below level of the Cranium.	18,000 ± 500
Asia Cemetery, Peru.	Dried vegetable matter, stringy and fibrous in appearance, from site 60/1 Asia cemetery. Tomb 21. Fill from false burial.	3,270 ± 100
Rio Seco del Leon, 87 km north of Lima, Peru.	Leaves and matting from excavations of house of pre-ceramic era. Site 52B. The leaves were covered with reed matting; the matting itself was sealed with clay and buried under stone boulders.	3,740 ± 100
	Charcoal from lowest level (200-250 cm) of Mound I—Site 52B.	3,800 ± 100
Las Aldas—330 km. north of Lima, Peru.	Mixture of carbonized wood, twiglets, pieces of bark, moss, grass Site 17/20.	
	Charcoal from Level No. 1—Chavinoid.	2,500 ± 100
	Charcoal from Level No. 3—pre-ceramic.	3,800 ± 80

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
213 214	R 370/3 R 370/4	F. Engel, Abancay, 377, Lima, Peru		
215	R 327/1	E. J. Searle, Auckland University, Auckland, N.Z.	N42/273427	N42/538
216	R 256	E. J. Searle, Auckland University, Auckland, N.Z.	N42/277564	N42/534
217	R 183/2	E. J. Searle, Auckland University, Auckland, N.Z.	N42/382539	N42/525
218 219	R 286/1 R 286/2	E. J. Searle, Auckland University, Auckland, N.Z.	N42/282679	N42/535 N42/536
220 221	R 326/1 R 326/2	R. N. Brothers, Auckland University, Auckland, N.Z.	N38/421726	N38/519 N38/520
222	R 326/3	R. N. Brothers, Auckland University, Auckland, N.Z.	N38/423728	N38/521
223	R 56/1	W. E. Begbie, Mount Albert Borough Council, Auckland, N.Z.	N42/225565	N42/520
224	R 183/3	E. J. Searle, Auckland University, Auckland, N.Z.	N42/266657	N42/526

Locality	Sample Details	Age Years before 1950
Otuma, Lagons de O'Iono—280 km. south of Lima, Peru.	Partially carbonized twiglets, branches, grass, and fibre. Site 104 FE. Charcoal from cut 12—level 1.	3,600 ± 80
	Charcoal from bottom of cut 12A1—level 1.	3,850 ± 80
South shore of Ihumatao, Mangere, Auckland, N.Z.	Wood from outer portion of tree in position of growth, enclosed by tuff forming a 15 ft cliff. These bedded tuffs are covered with flows from Ihumatao. The tree is part of an extensive fossil forest exposed on the shore platform.	29,000 ± 1,500
Woodside Road, Mt. Eden, Auckland, N.Z.	Wood, covered with ash, from sewage tunnel, near the margin of the Mt. Eden lavas. Ash probably from the Three Kings and sample probably antedates Mt. Eden basalt.	28,000 ± 1,000
Tamaki River, Auckland, N.Z.	Peat, underlying Panmure basin tuff. This sample will set a lower age limit for volcanic activity at this centre and may date a constructional terrace.	28,000 ± 1,000
Cnr. Anzac and Tauhoroto Sts, Takapuna, Auckland, N.Z.	Wood, directly overlain by basalt and tuff from Pupuke volcanic eruptions.	> 42,000
	Peat, covering a low terrace on shore of Shoal Bay, now partly capped by volcanics from Lake Pupuke Volcano.	> 40,000
West side of Administration Bay, Motutapu Island, Auckland, N.Z.	Shells from ash-free sand underlying the Rangitoto ash. Will give a maximum date for the Rangitoto eruptions.	750 ± 50
	Carbon sample from base of an oven overlying the 8 ft thick Rangitoto ash. Will give a minimum date for the Rangitoto eruptions and will also date the Moa-hunter flake culture at this level.	280 ± 40
East side of Administration Bay, Motutapu Island, Auckland, N.Z.	Wood, carbonized, underlying Rangitoto ash and overlying old greywacke soil. Will give actual age of Rangitoto ash shower.	770 ± 50
Oakley Creek Quarry, Mount Albert, Auckland, N.Z.	Charred branch beneath 15 ft of sub-recent basalt lava.	> 30,000
Bore No. 3 Harbour Bridge, Auckland, N.Z.	Carbonaceous soil underlying 55 ft of tuff, west of Onepoto Lagoon tuff ring. Will set a lower age limit for volcanic activity at this centre.	> 42,000

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
225	R 56/5	J. A. Bartrum, Auckland University, Auckland, N.Z.	N42/335530	N42/524
226	R 56/3	E. J. Searle, Auckland University, Auckland, N.Z.	N42/295688	N42/522
227	R 56/4	E. J. Searle, Auckland University, Auckland, N.Z.	N42/277686	N42/523
228 229	R 194/1A R 194/1B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
230	R 194/2A	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
231 232	R 194/3A R 194/3B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
233 234	R 194/4A R 194/4B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
235 236	R 194/5A R 194/5B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
237 238	R 194/12A R 194/12B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
239 240	R 232/13A R 232/13B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
241 242	R 232/3A R 232/3B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
243 244	R 279A R 279B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
245 246	R 232/5A R 232/5B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		

Locality	Sample Details	Age Years before 1950
Mt. Short, Penrose, Auckland, N.Z.	Charcoal, chipped out of basalt block from Mt. Short flow.	9,000 ± 160
Foreshore between Takapuna Beach, and O'Neil's Road.	Wood, enclosed in basalt flow.	Modern < 200
Smales Quarry, Takapuna, Auckland, N.Z.	Charcoal from cinders in east face of quarry.	> 36,000
Santa Barbara Basin, California, U.S.A.	Sediments 0-7·5 in. depth, core 4701	
	Carbonate fraction	2,260 ± 250
	Organic carbon fraction	3,000 ± 700
Santa Barbara Basin, California, U.S.A.	Sediments 165-170 in. depth, core 4701	
	Carbonate fraction.	5,600 ± 180
Catalina Basin, California, U.S.A.	Sediments 1-5 in. depth, core 4704	
	Carbonate fraction.	2,320 ± 130
	Organic carbon fraction.	1,970 ± 150
Catalina Basin, California, U.S.A.	Sediments 5-10 in. depth, core 4704	
	Carbonate fraction.	4,270 ± 130
	Organic carbon fraction.	2,820 ± 140
Catalina Basin, California, U.S.A.	Sediments 160-165 in. depth, core 4704	
	Carbonate fraction.	23,100 ± 1,000
	Organic carbon fraction.	18,400 ± 600
San Nicolas Basin, California, U.S.A.	Sediments 0-2 in. depth, core 4693	
	Carbonate fraction.	3,100 ± 100
	Organic carbon fraction.	2,560 ± 110
San Nicolas Basin, California, U.S.A.	Sediments 130-135 in. depth, core 4693	
	Carbonate fraction.	18,600 ± 400
	Organic carbon fraction.	16,900 ± 350
San Diego Trough, California, U.S.A.	Sediments 135-140 in. depth, core 4667	
	Carbonate fraction.	27,300 ± 1,500
	Organic carbon fraction.	21,900 ± 700
California, U.S.A.	Sediments 152-156 in. depth, core 4647	
	Carbonate fraction.	6,260 ± 250
	Organic carbon fraction.	4,900 ± 140
San Clemente Basin, California, U.S.A.	Sediments 120-125 in. depth, core 4670	
	Carbonate fraction.	12,400 ± 300
	Organic carbon fraction.	12,500 ± 250

N.Z. ^{14}C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
247 248	R 232/7A R 232/7B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
249	R 232/18B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
250	R 232/19B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
251	R 232/23A & B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
252	R 232/9B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
253	R 232/17B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
254	R 334/2B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
255	R 232/11B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
256	R 232/8B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
257	R 232/14B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
258	R 334/1B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
259	R 232/21B	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
260	R 232/6	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		

Locality	Sample Details	Age. Years before 1950
East Cortes Basin, California, U.S.A.	Sediments 173-178 in. depth, core 4671 Carbonate fraction. Organic carbon fraction.	 21,200 ± 500 21,000 ± 550
Long Basin, California, U.S.A.	Sediments, ocean floor. Core 4699, 135-140 in. depth.	13,600 ± 500
Catalina Basin, N.W., California, U.S.A.	Sediments, ocean floor. Core 4703, 105-110 in. depth.	8,250 ± 200
Deep Sea, California, U.S.A.	Sediments, ocean floor. Core 4698, 40-44 in. depth.	17,100 ± 1,100
No Name Basin, California, U.S.A.	Sediments, ocean floor. Core 4673, 158-163 in. depth.	32,600 ± 1,700
Continental Slope, California, U.S.A.	Sediments, ocean floor. Core 4697, 150-156 in. depth.	26,100 ± 1,000
California, U.S.A.	Sediments, ocean floor. Core 4700, 135-140 in. depth.	9,000 ± 200
West Cortes Basin, California, U.S.A.	Sediments, ocean floor. Core 4674, 165-170 in. depth.	32,800 ± 1,300
San Clement Basin, California, U.S.A.	Sediments, ocean floor. Core 4672, 105-110 in. depth.	11,800 ± 300
Tanner Basin, California, U.S.A.	Sediments, ocean floor. Core 4696, 130-135 in. depth.	20,200 ± 600
California, U.S.A.	Sediments, ocean floor. Core 4679, 118-122 in. depth.	9,450 ± 250
San Pedro Basin, California, U.S.A.	Sediments, ocean floor. Core 4690, 130-137 in. depth.	7,050 ± 120
East Cortez, Basin, California, U.S.A.	Sediments, ocean floor. Core 4671, 2-5 in. depth.	3,250 ± 300

N.Z. 14C No.	Lab. No.	Collector	Sheet and Grid Ref.	N.Z. Fossil No.
261	R 232/20	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
262	R 232/10	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
263	R 365/1	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		
264	R 365/2	W. L. Orr and K. O. Emery, University of Southern California, Los Angeles, U.S.A.		

Locality	Sample Details	Age Years before 1950
San Pedro Basin, California, U.S.A.	Sediments, ocean floor. Core 4690, 2-10 in. depth.	2,500 ± 150
West Cortez Basin, California, U.S.A.	Sediments, ocean floor. Core 4674, 1-5 in. depth.	4,800 ± 200
Santa Barbara Basin, California, U.S.A.	Sediments, ocean floor. Core 4701, 160-165 in. depth.	5,150 ± 180
Santa Barbara Basin, California, U.S.A.	Sediments, ocean floor. Core 4701, 90-95 in. depth.	3,500 ± 150

ERRATUM

N.Z. ^{14}C No. 27 (*N.Z. J. Sci. Tech. B38* : 734)
for N.Z. Fossil No. S84/504
read S84/506

A PLIOCENE WHALE-BARNACLE FROM HAWKE'S BAY, NEW ZEALAND

By C. A. FLEMING, New Zealand Geological Survey, Department of Scientific and Industrial Research

(*Received for publication, 26 November 1958*)

Summary

An incomplete compartment of a whale-barnacle from the Waitotaran Stage of Waihua River, Hawke's Bay, is described as *Coronula aotea* n.sp. It is considered to be a close relative of the living species, *Coronula diadema* (Linnaeus).

INTRODUCTION

In a survey of the Tertiary history of the Cirripedia, Withers (1953, p. 96) recorded the genus *Coronula*, comprising the whale-barnacles, only from Pliocene and younger strata. Fossil records of *Coronula* are known from the Red Crag of Suffolk, England (Darwin, 1854b), from the Astian and Pleistocene of Italy (Alessandri, 1906), from the Neogene of Japan (Hatai, 1939) and Formosa (Hayasaka, 1934), from the Tertiary of Chile (Schneider, 1936), and from the Pliocene of Equador (Pilsbry and Olsson, 1951).

Whale-barnacles have a potential value for correlation of Late Tertiary sediments in distant parts of the world. The modern species are widely distributed by the pelagic whales on which they live, and there is no reason to doubt that their Late Tertiary ancestors were equally widespread. On the other hand, they are rare as fossils and too little is yet known of the stratigraphic distribution of the several extinct species for their use in correlation.

The specimen described below, the first fossil *Coronula* to be recorded from New Zealand, was collected by Mr R. Stoneley, of B.P. Exploration Company, during work for B.P. Shell and Todd Petroleum Development Limited, and brought to the attention of the author by Dr J. Marwick, Consulting Paleontologist to the company.

N.Z. J. Geol. Geophys. 2 : 242-7.

Subclass CIRRIPEDIA

Order THORACICA

Suborder BALANOMORPHA

Family BALANIDAE

Subfamily CORONULINAE

Genus ***Coronula*** Lamarck

1803 *Annal. Mus. 1* (6):464

Type species (Pilsbry, 1916): *Balanus diadema* Linnaeus

Coronula aotea n.sp.

(Figs 1–10)

HOLOTYPE: An incomplete compartment, presented to the New Zealand Geological Survey by B.P. Exploration Company. Height 24·5 mm.

DESCRIPTION: The compartment most closely resembles (and is described as) a left lateral or carinolateral (Fig. 1), but could perhaps be a rostral lacking the right radius and adjacent radial fold of the paries.

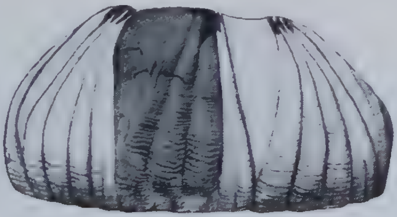


FIG. 1.—*Coronula aotea* n.sp. Diagram showing holotype compartment and reconstruction. × 1.

Viewed externally (Figs 1, 2), the compartment consists of a smooth radius attached to three sculptured longitudinal ribs, sub-equal in width, which correspond with the transverse loops of the folded walls forming two of the three basal cavities, and the half-fold bordering the radius; the fourth fold of the compartment, corresponding with the third transverse loop and basal cavity, has been lost.

Internally (Fig. 5), the specimen shows the basal ends of three sets of tranverse loops, attached to three septa (folded walls) forming two basal cavities. Above, the two folded walls on the rostral side have been broken off, but the third folded wall, on the carinal side, runs up to join the triangular special plate on which the alar rests.

The specimen is obviously closely related to *C. diadema* (Linnaeus), which it resembles in the external sculpture of the parietal ribs and in the unbranched ribs on the carinal side of the suture (Darwin, 1854a, p. 148). It also agrees with *C. diadema* in the delicate detail of sculpture of the double-walled partitions separating the basal cavities, and of the basal ends of the transverse folds of the wall. It has been compared with a series of living *C. diadema* from New Zealand and shows the following differences:

(1) Whale-barnacles vary greatly in shape and symmetry owing to crowding and accidents in growth, so that it is hazardous to draw conclusions on the shape of the whole shell from the profile of a single compartment. The contour of the holotype of *C. aotea* (Figs 3, 7) suggests, however, that the shell differed significantly from *C. diadema* in form. If the shell was crown-shaped, as in *C. diadema*, with more or less vertical walls (Fig. 10), the basal cavity, occupied by the whale's skin, below the ends of the folded walls, must have been higher than in *C. diadema*. If, as seems more likely, the basal cavity was depressed, as in other species of *Coronula*, the shell must have been conical in shape (Fig. 9), more like the normal form in *C. reginae* and *C. complanata;* but, as a consequence, the body chamber must have been cylindrical (as in *C. dormitor* Pilsbry and Olsson) rather than cup-shaped (as in *C. diadema*).

(2) The half-fold bordering the radius is wider than in *C. diadema*, almost as wide as the main parietal folds, particularly at the base, and is clearly marked off from the adjoining smooth radius.

(3) The radius is finely vertically striated, but shows only the faintest sign of the transverse striations (accentuated by wear and corrosion) that are invariably present on the radii of *C. diadema*, especially towards their distal edges.

(4) The radius (Fig. 3) is narrower, in relation to the three remaining ribs of the parietes, than it usually is in *C. diadema;* this is a variable character, of doubtful systematic importance.

(5) At its sutural edge, the true radius is seen to be thicker, in relation to the complementary radius within, particularly towards the base, than in *C. diadema*, a relationship which probably accounts for the lack of transverse striations on its outer surface.

(6) The oblique denticulated septa of the sutural edge of the complementary radius do not persist on its internal surface as they do in *C. diadema*, but die out rather abruptly at the inner edge of the suture, so that the internal surface of the radius (i.e., the outer wall of the cavity occupied by ovaria, Darwin, 1854a, p. 406, pl. 16, f. 7) is practically smooth.

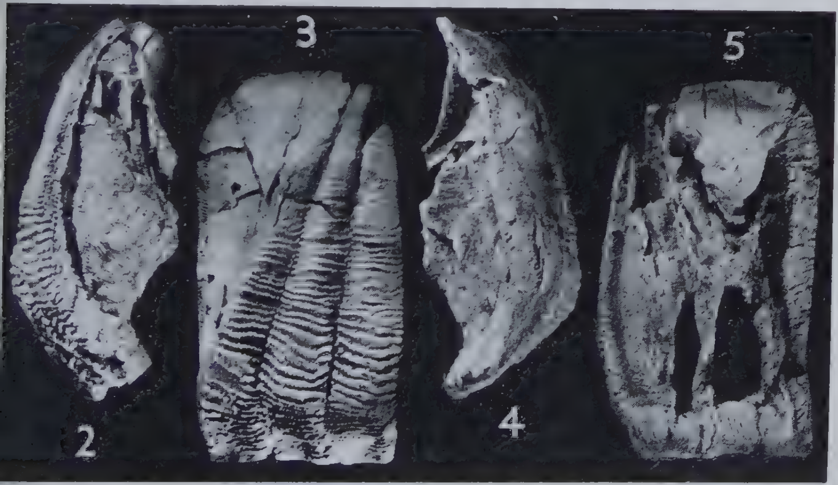
(7) The sutures between the ribs of the paries are minutely tuberculate, the tubercular teeth being larger and not so regularly arranged in transverse rows as in *C. diadema*. The tubercles are developed at the intersections of sinuous transverse elements with longitudinal striae, to give a somewhat cellular appearance to the suture (Fig. 2).

(8) The triangular special plate on which the ala rests (Fig. 5) is thick and solid, whereas in *C. diadema* its free edge is paper-thin and transparent. The rostral end of this plate bears traces of the denticles by which it is sutured to the ala; these are set on the same plane as the surface of the special plate, and not on a clearly defined step, as in *C. diadema*.

(9) The cavity between the radius and the special plates supporting the ala is neither solidly filled up, as in *C. barbara* (Darwin, 1854, pp. 38-39), nor open to the apex of the compartment, as in *C. diadema* and *C. reginae*, but is about two-thirds filled with solid shell, leaving a smoothly rounded, downward-facing depression (Fig. 7).

AFFINITIES: In the living species, *Coronula complanata* Mörch (= *C. balaenaris* auct.), the ribs branch symmetrically; the Italian Pliocene and Pleistocene species *C. bifida* Bronn (Alessandri, 1906) is similar (Pilsbry and Olsson, 1951, p. 203). These species form the subgenus *Cetopirus* Ranzani, recognized by Pilsbry (1916, p. 273).

The living species *Coronula reginae* Darwin and the fossil *C. ficarazzensis* De Greggorio (*fide* Pilsbry, 1916, p. 273) differ from *C. aotea* in having flat, more delicately sculptured parietal ribs, and a lower, conical, less cylindrical and crown-shaped shell.



FIGS 2 to 5.—*Coronula aotea* n.sp. Holotype, × 2.

(2) Rostral view, showing profile and sculpture of intercostal suture.

(3) Exterior, showing proportions of radius and parietal ribs.

(4) Carinal view, showing special plate on which the ala rests (indicating profile of body chamber) and sutural edge of radius.

(5) Internal view, showing special plate supporting the ala and basal ends of parietal ribs.

—*S. N. Beatus, photo.*

C. barbara Darwin 1854 (Red Crag of Suffolk; Pliocene of Orciano, Tuscany, Alessandri, 1906, p. 317) agrees with *C. aoteana* in the filling of the space between radii and alae, but differs in its coarser parietal sculpture and in details of the inner margin of the basis.

In *Coronula dormitor* Pilsbry and Olsson, 1951 (Pliocene, Equador), the parietal walls are more complexly folded, so that six basal cavities are present in a compartment of the same size as *C. aotea*. The apex of the compartment is not filled by shell.

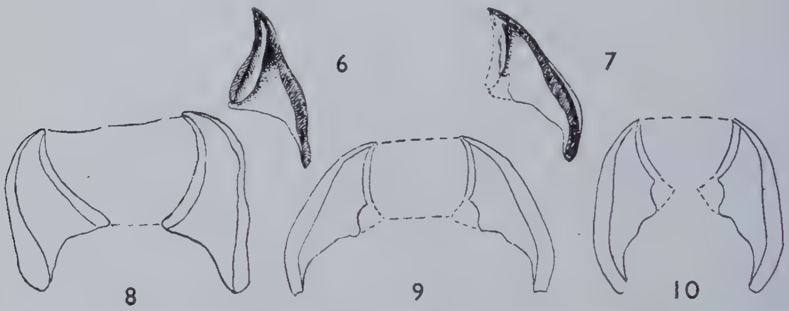


Fig. 6.—*Coronula diadema* (Linn.). Profile view (diagrammatic) of left lateral compartment (lacking ala) to show cavity running to apex of compartment. Based on a Recent specimen from New Zealand.

Fig. 7.—*Coronula aotea* n.sp. Profile view of holotype (diagrammatic) to show filling of cavity in apex of compartment.

Fig. 8.—*Coronula diadema* (Linn.) Cross-section of an asymmetric individual from Cook Strait (27 mm high).

Figs 9 and 10.—*Coronula aotea* n.sp. Hypothetical cross-sections based on holotype.

Two other fossil species of *Coronula* have been named. *C. antiqua* Philippi, 1887, Pleistocene (or younger) of Chile, is considered by Pilsbry and Olsson (1951, p. 203) to be a beach-rolled specimen of *C. diadema*; Schneider's more recent account of this form (1936) adds no diagnostic characters. *C. diluviana* Costa, 1834 (Naples), according to Pilsbry, appears to be either *C. bifida* or *C. barbara*.

From *Coronula diadema* (Linnaeus), *C. aotea* differs in the points enumerated above, of which the most important (*i.e.*, least liable to variation due to differences of environment and preservation) are the structure of the complementary alar plate and the partial infilling of the apex of the compartment, a feature that is of systematic importance in *C. barbara* Darwin. *C. aotea* may thus be characterized as a close relative of *C. diadema* that approaches *C. barbara* in this respect.

According to Mr H. E. P. Spencer, Ipswich Museum (pers. comm.), *C. barbara* occurs in the Newbournian (middle) zone of the Red Crag (Pleistocene), but it has also been recorded from the Pliocene (?Astian) of Italy (Alessandri, 1906). The living species *C. diadema* has been recorded as a fossil from the younger Tertiary of Formosa (Hayasaka, 1933–4) and from the Neogene of Japan (Hatai, 1939). The uncertain time relationships of these several fossil populations prevent any useful discussion of their phylogenetic relations.

LOCALITY AND AGE: B.P. Shell and Todd 8279, Waihua Valley, Hawke's Bay, N.Z.M.S.1, sheet N 115, grid reference 614954, coll. R. Stoneley, May 1958. Waitotaran.

The barnacle was associated with the Mollusca *Zelandiella allani* (King) and *Clavatoma pulchra* Powell. Two New Zealand Geological Survey collections from within 300 yards (GS 1560, 1565) include *Olivella neozelanica* Hutton, *Marshallena decens* Marwick, *Zeacuminia murdochi* Powell, and *Z. planitas* Laws. This association indicates that the beds are Lower Waitotaran (Waipipian Substage). The Waitotaran and underlying Opoitian Stage are currently correlated with the Astian–Plaisancian of Europe (Fleming, 1953), so *Coronula aotea* is approximately Upper Pliocene ("Astian").

ACKNOWLEDGEMENTS

The author is grateful to Mr N. L. Falcon, B.P. Exploration Co., London, for permission to record this interesting fossil. Dr R. K. Dell, Dominion Museum, supplied specimens of living *Coronula* for comparison. Mr J. Collins, London, and, through him, Mr H. E. P. Spencer, Ipswich Museum, sent data on European species. Mr H. O. Fletcher, the Australian Museum, kindly supplied a copy of a publication not available in New Zealand.

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AN ASSOCIATION OF HORNBLENDE AND CUMMINGTONITE FROM RINGARINGA, STEWART ISLAND, NEW ZEALAND

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Summary

Descriptions, with optical and chemical data, are given of hornblende and cummingtonite from an amphibolite exposed at the south end of Ringaringa beach, near Half Moon Bay, Stewart Island. The minerals are compared with similarly associated hornblende and cummingtonite described in recent years from Finland, and analyses are plotted to show the extent of the suggested miscibility gap between hornblendes and cummingtonites associated with them. The paragenesis of the hornblende and cummingtonite is briefly discussed.

A short apepndix describes the method used in separating the amphiboles from the crushed rock for analysis.

INTRODUCTION

The hornblende and cummingtonite to be described below occur in an amphibolite exposed at the south end of Ringaringa beach, near the settlement of Half Moon Bay, on the east side of Stewart Island. Although amphibolites, of igneous origin, are known from a number of localities on the north-east coast of the island,* the Ringaringa locality is the only one where cummingtonite has been found in appreciable amounts, forming one-third of the total amphibole content of the rock.† In this paper a brief description, with chemical and optical data, of the rock and its constituent minerals is given, and the paragenesis of the associated amphiboles discussed.

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*A general account of the geology of this area will be presented in a later paper.

†This is apparently the first record of cummingtonite-bearing rocks in place in New Zealand. Mason has recently (1953, 1955) described cummingtonite in schist boulders from several rivers on the west coast of the South Island.

N.Z. J. Geol. Geophys. 2 : 248-56.

PETROGRAPHY

The Ringaringa amphibolite forms large wave-planed outcrops that are cut in places by numerous veins of quartz-rich tonalite. Unfortunately, the surrounding area is not well exposed, and contacts of the amphibolite with possible older rocks are not visible. In hand specimen the amphibolite is a dense, rather fine-grained grey to dark grey rock showing on the broken surface a plexus of little feldspar prisms, with less abundant intervening crystals of amphibole. Under the microscope, the rock shows a fairly typical gabbroid texture (see Fig. 1); the grain-size is variable, but averages about $\frac{3}{4}$ mm. Plagioclase, lying within the acid to medium andesine range, forms 55 to 65% of the volume of the rock and occurs in sub-idiomorphic laths sometimes with distinct sub-parallel orientation. Albite twinning is present in nearly every crystal. Occasionally the cores of the larger crystals are clouded by dusty decomposition products that include a little sericite and clinozoisite. There are a few small interstitial crystals of quartz. Green hornblende (15 to 20%) occurs in irregular prismatic crystals occasionally bearing small ophitically included prisms of feldspar. Its optical properties are as follows: α 1·648, β 1·659, γ 1·668 (all $\pm$ 0·003); $\gamma - \alpha$ 0·020; $2V_{\alpha}$ 72°, 73°, 75°; $Z \wedge c$ 15°; pleochroic scheme, X light yellow-green, Y leaf-green to olive-green, Z blue-green.

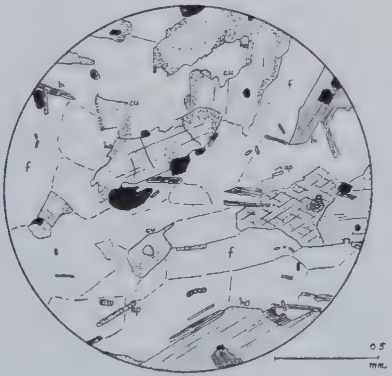

FIG. 1.—Camera-lucida sketch of cummingtonite-bearing amphibolite, Ringaringa, Stewart Island (ordinary light), f = feldspar, cu = cummingtonite, ho = hornblende, bi = biotite, ap = apatite.)

Throughout the rock the hornblende is intimately associated with a colourless or very pale green cummingtonite. The latter may form uniform discrete crystals, but usually it is zoned by a more or less continuous rim of green hornblende (Fig. 1). In crystals thus zoned the (110) cleavage traces are continuous between the two minerals. The line of demarcation between the amphiboles is usually quite distinct, but in some cases there seems to be a very thin graditional zone between the two. No case was seen of hornblende being rimmed by cummingtonite. The cummingtonite often shows fine striations at about 70° to the (110) cleavage traces. When present in zoned crystals, the striations are usually continued into the surrounding hornblende; rarely homogeneous crystals of hornblende may show the striations. Occasionally, the cummingtonite has multiple twinning parallel to (100). Optical properties are as follows: α 1·641, β 1·650, γ 1·667 (all ± 0·003); $\gamma - \alpha$ 0·026; $2V_{\gamma}$ 79°, 80°, 82°, 83°; $Z \wedge c$ 18°; disp. slight $r > v$.

Biotite (6 to 8%) occurs in long flakes and as small irregular crystals flecking the hornblende. Here and there it is altered to give thin lamellae of pale green chlorite. Its properties are: α 1·597, γ 1·651 (±0·003); $\gamma - \alpha$ 0·054; pleochroic scheme, X light straw-yellow, Y = Z dark chocolate-brown. Accessory minerals are magnetite grains and numerous tiny needles of apatite. A few small prisms of zircon were noted.

In Table 1 modes of the rock taken from three different thin sections are given:

TABLE 1.

Plagioclase (An_{32-34})	56·4	65·2	62·6
Quartz	0·9	1·5	1·0
Hornblende	30·2	16·5	18·6
Cummingtonite		6·3	5·0
Biotite	6·2	5·7	7·7
Magnetite	4·2	4·7	5·1
Apatite	2·1		

The analysis and norm of the rock are given below:

SiO_2	50·94
Al_2O_3	19·32
Fe_2O_3	3·80
FeO	6·63
MgO	3·49
CaO	7·66
Na_2O	4·84
K_2O	0·84
H_2O-	0·08
H_2O+	0·72
TiO_2	1·06
P_2O_5	0·61
MnO	0·22
	100·21

	or	5·00
	ab	40·87
	an	28·36
di	wo	2·55
	en	1·30
	fs	1·19
hy	en	2·20
	fs	1·85
ol	fo	3·64
	fa	3·47
	mt	5·57
	il	2·13
	ap	1·34

(Anal. W. A. Watters)

ANALYTICAL DATA

In the following table are given the analyses of the hornblende and cummingtonite separated from the Ringaringa amphibolite.*

	A				B			
		Metal atoms to 24 (O, OH)			(1)	(2)	Metal atoms to 24 (O, OH)	
SiO_2	45·88	6·789		8·000	51·68	52·05	7·726	7·862
Al_2O_3	8·83	1·528	1·211 0·317	5·177	1·73	0·77	0·136	
TiO_2	0·69	0·071			0·62	0·60	0·067	7·144
Fe_2O_3	4·45	0·497			3·55	3·40	0·380	
FeO	14·64	1·813			21·59	22·35	2·773	
MgO	10·90	2·399			14·85	15·26	3·374	
MnO	0·65	0·080			1·33	1·41	0·177	
CaO	10·92	1·733		2·070	2·98	1·89	0·300	
Na_2O	0·98	0·284			0·31	0·21	0·061	
K_2O	0·31	0·053			0·10	0·07	0·012	
H_2O+	2·13	2·097		2·097	2·02	1·98	1·96	1·96
	100·38				100·76	99·99		

Analyses of hornblende and cummingtonite separated from amphibolite at Ringaringa.

A. Hornblende.
B. (1) Cummingtonite with impurities of hornblende.
(2) Cummingtonite, re-calculated abter subtraction of impurities.

(Anal. W. A. Watters)

The structural formula ($X_{2·07}$ $Y_{5·177}$ Z_8O_{22} $(OH)_{2·097}$) calculated from the analysis of the hornblende corresponds reasonably closely to the standard amphibole formula. Nevertheless, the number of atoms (5·2) for the Y group (principally Mg and Fe^{++}) is somewhat higher than the ideal value of 5, whereas, on the other hand, the number of X group atoms (Ca, Na, K)—2·07—is less than the usual value for common green hornblendes. This feature has also been recorded by Seitsaari (1952, p. 7) for hornblende found in association with cummingtonite in other localities; its significance is not known, but it may be due to the presence of small amounts of an amphibole transitional to cummingtonite, a possibility that is suggested by Seitsaari (1952). The hornblende falls about half-way between edenite (ideally $NaCa_2Mg_5(Si_7AlO_{22})(OH)_2$) and pargasite ($NaCa_2Mg_4Al(Si_6Al_2O_{22})(OH)_2$) on the classification and diagram given by Sundius (1946, fig. 6, p. 19) for the lime-bearing amphiboles.

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*A note on the method of separation of the minerals is given as an appendix to this paper.

Due to the incomplete separation of the two amphiboles (see appendix), the structural formula ($X_{7\cdot144}Z_{7\cdot862}O_{22}(OH)_{1\cdot96}$) calculated from the analysis of the cummingtonite is not entirely satisfactory, the number of Z group atoms (mainly Si) being low. Nevertheless, the analysis is probably sufficiently close to the true value to allow a comparison with the associated hornblende to be made. The optical properties show satisfactory correspondence with the data recorded by Sundius (1931) and Bowen and Schairer (1935) for cummingtonites of similar composition, except that the optical axial angle appears to be a little high.

GRAPHICAL REPRESENTATION

Sundius (1933) has discussed fully the miscibility gaps existing between the various groups of amphiboles. He concluded that the pronounced gap between lime-bearing and associated lime-poor amphiboles lies between 2 and 23% $CaSiO_3$ (molecular) for the more magnesian members, and between about 5 and 23% for the more iron-rich members within the field $CaSiO_3$-$MgSiO_3$-$(Fe,Mn)SiO_3$). In fig. 2 the analyses of cummingtonite and associated green hornblende published by Eskola (1950) and Seitsaari (1952), together with those of the amphiboles from the Ringaringa rock, are plotted on the triangular diagram $CaSiO_3$-$MgSiO_3$-$(Fe,Mn)SiO_3$. It is believed that these analyses are the only available ones of conjugated hornblende and cummingtonite from the same rock. Because there is, however, an appreciable variation in the iron-magnesium ratio in the three sets of analysed amphiboles, the plotted analyses do not give a good indication of the extent of the gap between the two amphibole groups. The limits for the miscibility

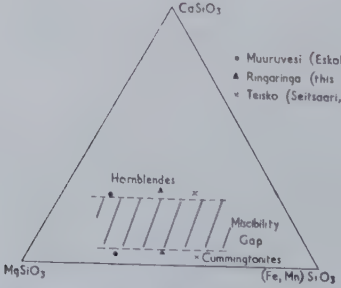

FIG. 2.—Triangular diagram showing the plots of three pairs of associtaed hornblende and cummingtonite within the field $CaSiO_3$-$MgSiO_3$-$(Fe,Mn)SiO_3$. The shaded area indicates the approximate extent of the suggested miscibility gap between the two groups of amphiboles.

defined by the analyses lie close to those drawn by Sundius (1933) and earlier by Asklund (1923, p. 30). Analytical data are not known for iron-rich conjugated members, and it is still unknown to what extent the gap is bridged by a possible intermediate amphibole that may occasionally form a thin transitional zone between hornblende and cummingtonite in zoned crystals (Seitsaari, 1952, p. 18; Kuno, 1950, p. 980).

PARAGENESIS

In the rock described above, zoned crystals always show cummingtonite mantled by hornblende. This texture is probably capable of more than one interpretation, and there is no definite evidence to show which of these minerals has crystallized first. To the writer, however, the development of hornblende rims around cores of cummingtonite rather than the reverse is more feasibly explained by assuming that the greater part of the hornblende has formed later than the cummingtonite during the evolution of the rock. This view is taken in the following discussion of the evolution of the two minerals.

The change from the cummingtonite to green hornblende requires the addition of alumina, lime, and minor soda to the cummingtonite molecule. One possible source of these components is anorthite. The relation between the amount of cummingtonite in the rock and the anorthite content of the plagioclase has been briefly discussed by several authors. For example, Stewart (1946, p. 483) showed that during uralitization of noritic rocks hypersthene may change to cummingtonite, which later reacts with anorthite to produce a green hornblende, thus: cummingtonite + anorthite → hornblende + silica (quartz). Asklund (1923, pp. 20–25) also recorded similar changes during the alteration of a noritic gabbro from Stavsjö, Sweden, the hornblende and quartz, produced as a result of this reaction, forming a myrmekite-like intergrowth. In the Ringaringa rock there is no petrographic evidence that the alumina, lime, and soda required in the change from cummingtonite to hornblende are now being derived from plagioclase, and the rock possibly represents a condition of equilibrium between the two amphiboles for the grade of metamorphism reached.

In some rock the association hornblende-cummingtonite forms part of a discontinuous reaction series, hypersthene → cummingtonite → horneblende (Asklund, 1923, p. 21; Nockolds, 1934a and 1934b, p. 311; Deer, 1935, p. 68; Iwao, 1937), the minerals apparently developing as a result of hybridism and reciprocal reaction between original basic igneous rocks—now represented by amphibolites—or calc-argillaceous hornfelses and invading acid (granitic or tonalitic) magma. Unfortunately, the nature of possible original ferromagnesian minerals in the Stewart Island rock is not known, but further field work may reveal the less altered parent rock and so provide more definite information on the paragenesis of the amphiboles.

Unlike the Ringaringa rock, certain amphibolites contain intergrowths of cummingtonite and hornblende that appear to indicate equilibrium between and apparent simultaneous crystallization of these two minerals

(Asklund, 1923, p. 27). Seitsaari (1952) shows that in the rock studied by him the crystallization of amphibole was evidently initiated sometimes by cummingtonite and sometimes by hornblende. According to Asklund (1923, pp. 20–32), two types of cummingtonite paragenesis in the alteration of gabbroic and noritic rocks may be recognized, without assuming introduction of material from outside. In rocks containing two pyroxenes the augite changes directly into green hornblende, whereas the hypersthene breaks down to cummingtonite which may later be replaced by hornblende (see also above). On the other hand, in some cummingtonite-bearing amphibolites the two amphiboles appear to be in equilibrium with one another and to have resulted from the breakdown of a single pyroxene. In these rocks the hornblende and cummingtonite may occur together as an intimate intergrowth.

Many amphibolitic rocks containing both hornblende and cummingtonite, for example, the Ringaringa rock and several of those described by Seitsaari (1952, pp. 10–16), show good preservation of original igneous textures, and, in fact, these rocks may sometimes be given typical igneous rock-names. Such rocks have probably not attained complete metamorphic equilibrium, and the cummingtonite present in them may well be a relict mineral that would have changed over completely to hornblende had the rock suffered further recrystallization and obliteration of its original texture. This problem would require further petrographic data before a final opinion on it could be stated.

ACKNOWLEDGEMENTS

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APPENDIX

METHOD OF SEPARATION OF HORNBLENDE AND CUMMINGTONITE FROM THE AMPHIBOLITE

For the mineral separation, the following procedure was used. The rock (about $1\frac{1}{2}$ lb) was crushed to pass a 150-mesh sieve and the dust washed off by decantation in water. After as much magnetite as possible was removed, the dried powder was separated into feldspar- and amphibole-rich fractions by passing it rapidly through the Frantz isodynamic separator. The feldspar-rich fraction was discarded. The dark, amphibole-rich fraction was passed slowly through the separator again. With an angle of tilt $= 27\frac{1}{2}°$, angle of slope $= 30°$, and with maximum current, there was a fairly good separation into two fractions, one consisting of hornblende with no cummingtonite but with occasional flakes of biotite and feldspar, the other containing about 50% cummingtonite with roughly equal amounts of hornblende and biotite and a little magnetite. The hornblende fraction, after repeated centrifuging in liquid of s.g. 3·1 to remove biotite and feldspar, was used for the analysis of the hornblende.

The second fraction was treated with a magnet to remove remaining magnetite and mineral grains with magnetite inclusions. Biotite was then removed by centrifuging in liquid of s.g. 3·1. Separation of the cummingtonite and hornblende making up the remaining powder was first attempted by centrifuging in methylene iodide-bromoform mixtures, but without success. On the isodynamic separator, trial and error showed that with angles of tilt and slope = 10°, and with medium current (0·8 amp.), a fairly good, though still incomplete, separation was effected. The powder was passed through the separator very slowly several times. After each run the hornblende was rejected and the cummingtonite-rich fraction passed through again. The small amount (1·6 g) used for analysis still contained about 10% of hornblende, the amount of which was carefully determined by counting the grains in several fractions with a graduated eyepiece.

HOCHSTETTER CENTENARY ISSUE

The year 1959 is the centenary of the field work in New Zealand of Ferdinand von Hochstetter, geologist attached to the expedition round the world of the Austrian frigate *Novara* in 1857-1859. Hochstetter's geology was not the result of casual observations; at the invitation of the Provincial Governments of Auckland and Nelson he explored a large part of south Auckland Province from Auckland isthmus south to Taupo, and the more accessible parts of Nelson, as a basis for topographic and geological maps. His final report, incorporating all that was then known on the subject, was published in 1864 under the title "Geologie von Neu-Seelande". Hochstetter's work was the first major contribution to the geology of this country, and established a tradition of regional geological mapping associated with careful paleontological and stratigraphic studies that does much to explain why New Zealand is so far ahead of many larger countries in fundamental geological mapping.

It is proposed to issue a special Hochstetter Centenary number of the *New Zealand Journal of Geology and Geophysic* towards the end of 1959, to include contributions on the geology of areas investigated by Hochstetter, historical accounts, critical evaluations of his work and of the fossil plants and animals described from his collections. Manuscripts and illustrations for this issue should be submitted to the Editor, P.O. Box 8018, Wellington, as early as possible in 1959.

N.Z. J. Geol. Geophys. 2 : 257.